Southeastern Geology: Volume 19, No. 2 February 1978

Edited by: S. Duncan Heron, Jr.

Abstract

Academic journal published quarterly by the Department of Geology, Duke University.

SOUTHEASTERN GEOLOGY

PUBLISHED AT DUKE UNIVERSITY DURHAM, NORTH CAROLINA

VOL. 19 NO. 2 FEBRUARY, 1978

SOUTHEASTERN GEOLOGY

PUBLISHED QUARTERLY

AT

DUKE UNIVERSITY

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SOUTHEASTERN GEOLOGY

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UPPER COASTAL PLAIN SURFICIAL SEDIMENTS BETWEEN THE TAR AND CAPE FEAR RIVERS, NORTH CAROLINA[1]

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ABSTRACT

The Cretaceous Middendorf and Pliocene Pinehurst Formations are the major surficial sediments on the Neuse-Cape Fear drainage divide in the upper Coastal Plain (> 275 feet) of North Carolina. The Miocene Macks Formation is exposed on valley sides in limited areas along the eastern end of its distribution, but it is not a major surface deposit. The fluvial member (i. e., not marine or eolian) of the Pinehurst, throughout most of its areal distribution, is west of the Coats scarp at surface altitudes above 275 feet. In most areas south of the Neuse River, the Coats scarp and its subsurface equivalent truncates the Macks and Pinehurst Formations. Near Bailey, North Carolina, the Coats scarp is an erosional feature formed entirely on the Pinehurst Formation. In this area the Pinehurst extends eastward to the Wilson Mills scarp as the surficial sediment to a surface altitude of about 230 feet. This interpretation means that the Brandywine geomorphic surface at and north of Bailey, North Carolina, is an erosional surface on the Pinehurst Formation, whereas south of Bailey, it is a depositional surface related to an unnamed post-Pinehurst Formation.

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[1]Joint contribution from the Soil Conservation Service, USDA, and the Soil Science Department, North Carolina Agricultural Experiment Station. Paper No. 5131 of the Journal Series of the North Carolina Agricultural Experiment Station, Raleigh, North Carolina 27607.

INTRODUCTION

The surficial sediments of the upper Coastal Plain, the sediments above the Coats scarp, have been investigated in varying degrees since about 1890. McGee (1890) called these deposits the Appomattox Formation (Table 1) but he included sediments at elevations as low as 25 feet. Johnson (1907) recognized a terrace level that he considered to be Pleistocene between 280 and 320 feet. Stephenson (1912) modified the earlier work of McGee and Johnson when he renamed part of the Appomattox the Lafayette Formation and placed its eastern boundary at a discontinuous scarp with a toe altitude of about 320 feet. The Lafayette Formation had to be abandoned almost before it was proposed by Stephenson because Berry (1911) found that the type section of the Lafayette in Mississippi was Eocene, not Pliocene as originally thought. Cooke (1930) later suggested that the Lafayette mapped by Stephenson in North Carolina included a terrace formation that had a shore line at about 265 feet. He believed this unit was the correlative of the typical Brandywine Formation in Maryland and the Hazlehurst Formation of Georgia.

The post-Cretaceous upper Coastal Plain deposits were more or less ignored until Mundorf (1946) and Richards (1950) gave them the informal name "high level gravels." Doering (1960) made regional correlations from Alabama to Virginia and renamed these deposits the Citronelle Formation, but his idea of the Citronelle was much like McGee's Appomattox because it extended eastward to include Cooke's Brandywine and Coharie Formations. In 1962, Conley renamed these high Coastal Plain deposits in southern North Carolina (Moore County) the Pinehurst Formation and emphasized the gravelly and sandy nature of these deposits capping the stream divides. Daniels et al. (1966) applied the name Pinehurst to a fluvial sequence of sand and gravels grading upward into finer textured sediments near Benson, North Carolina. Their Pinehurst was confined to the upper Coastal Plain, i. e., above the Coats scarp that has a toe altitude of about 275 feet. The Pinehurst near Benson appeared to be very close to Conley's idea of the formation and inspection of several road cuts in Moore County indicated a great similarity between these sediments. Cooley (1970) restudied the type section of Conley's Pinehurst and found it to be of shoreline marine and eolian origin. Conley recognized only the Pinehurst at his type section, but Cooley has shown that the lower part of the exposure is a gravelly member of the Middendorf Formation. Bartlet (1967) questioned the validity of the fluvial origin of the Pinehurst when he mapped the formation as an eolian unit in the Southern Pines 15 minute qudrangle.

Despite the questions about the Pinehurst raised by later workers, we consider this formation name to be the best of several alternatives. New formation names at this time seem inappropriate because so little detailed local or regional work has been done, and using names, such as the Citronelle, that have been applied to several deposits only adds to the confusion. We therefore use the term "fluvial member" of

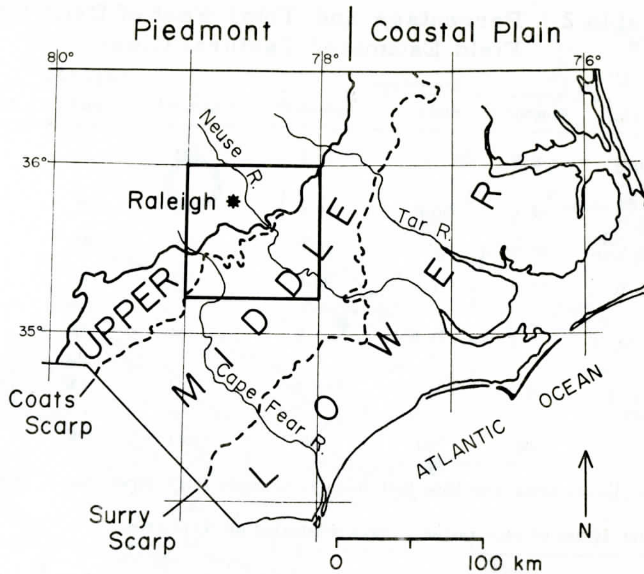

Figure 1. Location map showing the upper, middle and lower divisions of the Coastal Plain. The area of this study is in the square shown with heavy lines.

Table 1. Correlation of Upper Coastal Plain Deposits.

	McGee 1890	Johnson 1907	Stephenson 1912	Mundorf 1946	Richards 1950	Doering 1960	Conley 1962	Daniels et al 1966	Cooley 1970	This paper
Pleistocene		?				Citronelle				
Pliocene	Appomattox		Lafayette	High level gravels	High level gravels		Pinehurst	Pinehurst	Pinehurst	Pinehurst Fluvial member
Miocene								Macks	--	Macks
Cretaceous					Tuscaloosa		Tuscaloosa	Tuscaloosa	Midden-dorf	Middendorf

the Pinehurst for those post-Cretaceous deposits above 275 feet that are not marine or eolian in origin.

Our area of investigation is between the Tar and Cape Fear Rivers (Figure 1) although some work has been done south of the Cape Fear River to help correlate our work with others. Most of the investigations have been made by auger drill rig, partly because most road cuts in the area do not penetrate to the least weathered parts of the surficial units. When possible, the drill stem was rotated into the sediments at the pitch of the auger lands and pulled, without further rotation, every 5 to 10 feet to minimize disturbance of the beds on the auger flights.

Table 2. Percentage and Total Feet of Drill Hole in Each Field Estimated Textural Class.

Textural Class[1]	Percentage			Total Feet[2]		
	Pinehurst	Macks	Middendorf	Pinehurst	Macks	Middendorf
Sand to loamy sand	6.4	-	10.6	133	--	78
Sandy loam	39.1	40.8	39.6	806	301	292
Sandy clay loam	40.9	11.5	14.1	839	85	104
Sandy clay to clay	11.7	17.9	31.9	242	115	234
Clay loam to loam	1.5	21.6	1.8	34	176	14
Silty clay to silt loam	0.3	8.2	2.0	7	61	15
TOTALS	100	100	100	2,061	738	737

[1]Textural classes based upon USDA Soil Textural Triangle (Soil Survey Staff, 1951, p. 209).

[2]Total feet of bed of each textural class penetrated by drill stem.

STRATIGRAPHIC SEQUENCE

The unconsolidated sediments in the upper Coastal Plain are the Middendorf, Macks, and Pinehurst Formations. The Middendorf, called Tuscaloosa by earlier workers (Stuckey and Conrad, 1958, Daniels et al., 1966) has an extremely variable lithology, ranging from tough white or gray clays to coarse micaceous sands. About 40 percent of the Middendorf is a sandy loam and about 32 percent clay or sandy clay (Table 2). Kaolinite is the dominant clay mineral in the Middendorf (Heron, 1960), although near Benson, North Carolina, when the Middendorf interfingers with the Black Creek Formation, the darker massive clays of the Middendorf are rich in montmorillonite (Craig, et al., 1972; Daniels and Gamble, in preparation). Abrupt vertical and horizontal changes in lithology are common. Silicified logs have been found in the sandy beds but we have found no other fossils in typical Middendorf beds. An excellent exposure of Middendorf is the Durham and Southern Railroad cut between Holland and Fuquay Varina. The cut is at the railroad underpass to Wake County road 1107 near its junction with road 2768. Almost all the recognizable Middendorf is south of the Neuse River (Figure 2).

The Macks Formation (Daniels et al., 1966) disconformably overlies the Middendorf south of the Neuse River. The contact to the underlying sediments is sharp, commonly marked by pebbles, and is smooth with a local relief of about 5 feet near Benson. North of the Neuse River, the Macks rests upon saprolite (Figure 3, 6) and the relief at the base may be considerably more than 5 feet. The Macks Formation is a very distinct lithologic unit that south of the Neuse River is

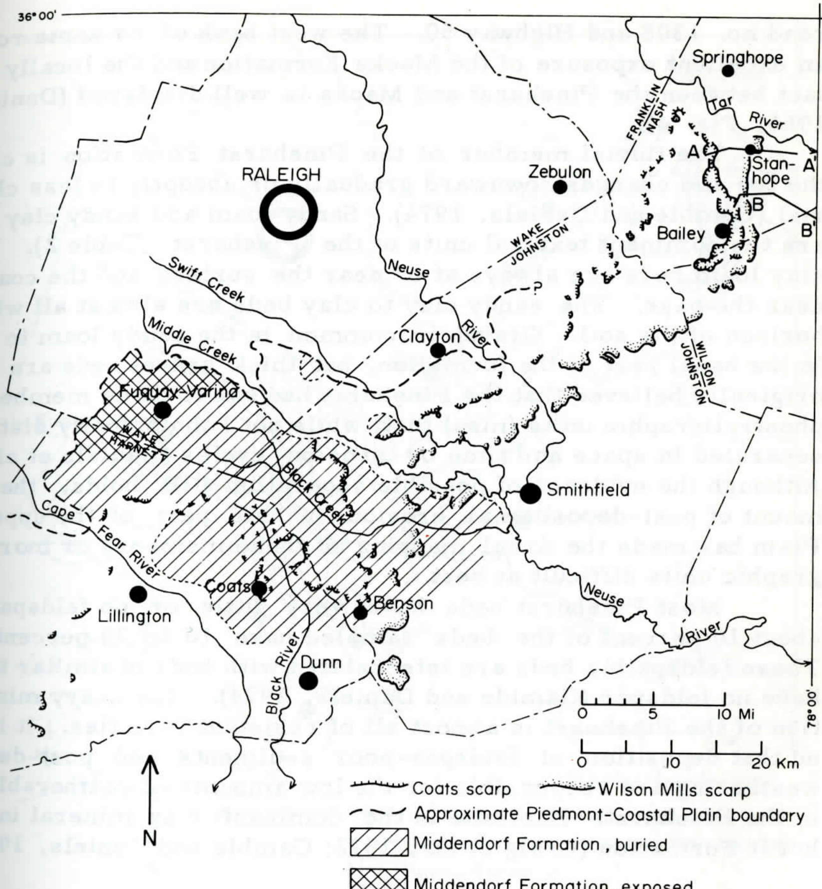

Figure 2. Distribution of Middendorf Formation.

a yellowish brown to gray and sometimes red, massive fine to very fine sandy loam to sandy clay loam with about 52 percent of the formation in these two textural classes (Table 2). Clayey beds are not common but clay loam to loam beds containing considerable silt make up about 22 percent of the formation. South and east from Bailey (Figure 3), the Macks is red (2.5 YR 4/6) on exposed outcrops instead of the more common yellowish brown of the exposures south of the Neuse River.

The fluvial member of the Pinehurst overlies saprolite and the Middendorf, and Macks Formations in the study area (Figures 4, 5, 6). The contact is an erosion surface commonly marked by gravel with a relief of 15 feet or more (Daniels et al., 1966). Most sections of the Pinehurst on interstream divides have soils 10 feet or more thick (Daniels et al., 1970), and bedding in most road cuts is poorly preserved. One of the better exposures displaying some bedding is on the east side of the NC Highway 50 just north of the junction of Johnston County road no. 1308 and Highway 50. The west bank of the same road cut has an excellent exposure of the Macks Formation and the locally rough contact between the Pinehurst and Macks is well displayed (Daniels et al., 1966, Fig. 4).

The fluvial member of the Pinehurst Formation is clayey near the top and changes downward gradually or abruptly to less clayey material (Gamble and Daniels, 1974). Sandy loam and sandy clay loam beds are the dominant textural units of the Pinehurst (Table 2). The sandy clay loam beds are always at or near the surface and the coarser beds near the base. The sandy clay to clay beds are almost all within the B horizon of the soil. Gravel is common in the sandy loam to sand beds in the basal part of the formation, but thick gravel beds are rare. We originally believed that the Pinehurst had two to three members or morphostratigraphic units (msu) that, while not lithologically distinct, were separated in space and time by erosion cycles (Daniels et al., 1966). Although the evidence of possible separation still holds, the large amount of post-depositional erosion in this part of the upper Coastal Plain has made the areal mapping of these members or morphostratigraphic units difficult at best.

Most Pinehurst beds have very little or no feldspar although about 10 percent of the beds sampled have 10 to 20 percent feldspar. These feldspathic beds are intercalated with beds of similar texture that have no feldspar (Gamble and Daniels, 1974). The heavy mineral fraction of the Pinehurst is almost all of resistant varieties. It is suggested that deposition of feldspar-poor sediments and post-depositional weathering are responsible for the low amounts of weatherable minerals in the Pinehurst. Kaolinite is the dominant clay mineral in the Pinehurst Formation (Craig et al., 1972; Gamble and Daniels, 1974).

Distribution of Formations Along Interstream Divides

The Middendorf Formation has not been identified in subcrop north of the Neuse River (Figure 2). South of the Neuse River, the Middendorf is continuous under the divides, and it is exposed at several areas in topographic lows and in one large area northwest from Fuquay near Wilbon above 450 feet (Figure 5). The Middendorf is exposed on the valley sides of the Black Creek valley near Route 50 (Daniels et al., 1966, Fig. 2). The Macks Formation, while discontinuous under the Neuse-Cape Fear divide, is exposed on the valley slopes of Black Creek. The Macks is seldom exposed on the interstream divides (Figures 5 and 6), and northwest from Route 210 on the Neuse Cape Fear Divides it occurs as small discontinuous bodies to altitudes of about 420 feet (Figure 5). The Pinehurst Formation is nearly continuous, except in topographic lows or saddles, along the Neuse Cape Fear Divide from the toe of the Coats scarp at about 275 feet to altitudes of 420 to 450 feet (Figure 5).

Producing an accurate map showing the outcrop area of the Pinehurst and Middendorf Formations where the Macks Formation is absent

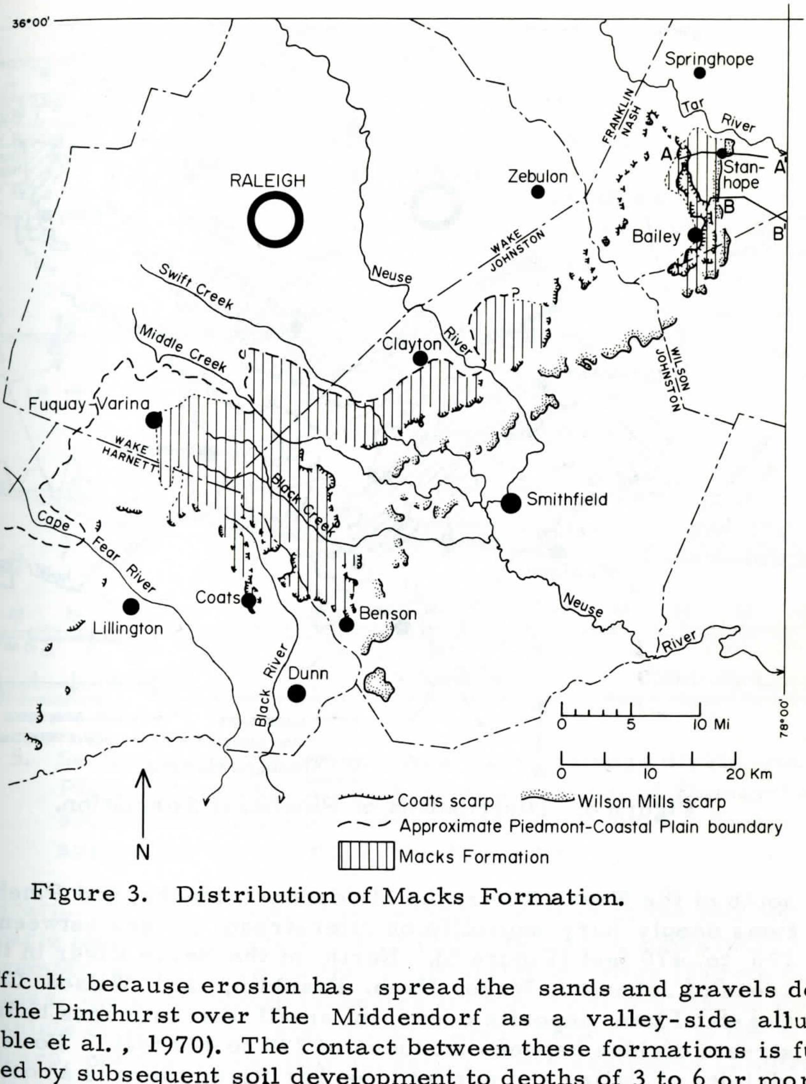

Figure 3. Distribution of Macks Formation.

is difficult because erosion has spread the sands and gravels derived from the Pinehurst over the Middendorf as a valley-side alluvium (Gamble et al., 1970). The contact between these formations is further masked by subsequent soil development to depths of 3 to 6 or more feet. For these reasons, we do not show a detailed outcrop map of the Pinehurst, Macks, and Middendorf Formations.

South of the Neuse River, the eastward limit of the Pinehurst Formation is the Coats scarp with a toe altitude of about 275 feet (Figure 4). In this area, the erosion surface at the base of the Brandywine morphostratigraphic unit (msu) truncates the Macks and Pinehurst Formations and rises to the surface as the Coats scarp (Daniels et al., 1966; 1972). The characteristics of the Brandywine and Coharie have been described elsewhere (Daniels and Gamble, 1974).

North of the Neuse River the stratigraphic and geomorphic situation appears to be completely different. In Wake and Johnston Counties

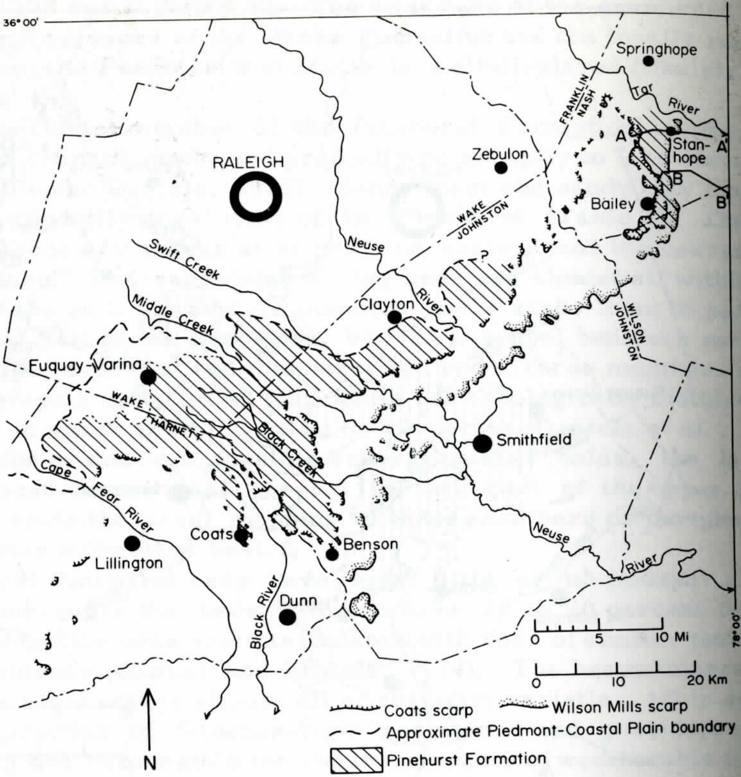

Figure 4. Distribution of Pinehurst Formation.

south of the Neuse River, the Middendorf, Macks, and Pinehurst Formations deeply bury saprolite on interstream divides between altitudes of 270 to 470 feet (Figure 5). North of the Neuse River in the dissected areas of Johnston, Wake, Nash, Franklin, and Wilson Counties, the Coastal Plain deposits occur in small discontinuous bodies on interstream divides between altitudes of 270 to 470 feet. However, in the vicinity of Bailey and north to the Tar River there is a continuous cover of Coastal Plain deposits up to altitudes of 300 feet or more. Most of the correlation with deposits south of the Neuse River will be from the Bailey area.

Two traverses, one along North Carolina Route 97 near Stanhope and the other near the settlement of Mt. Pleasant in Nash County (Figure 6) give a completely different relation between sedimentary bodies and the major Coastal Plain scarps than that seen south of the Neuse. These traverses indicate that the Pinehurst Formation north of the Neuse River extends eastward from the Coastal Plain-Piedmont contact at altitudes about 300 feet to altitudes of about 230-210 feet. We are unable to find any subsurface break in sediment that correlates with the

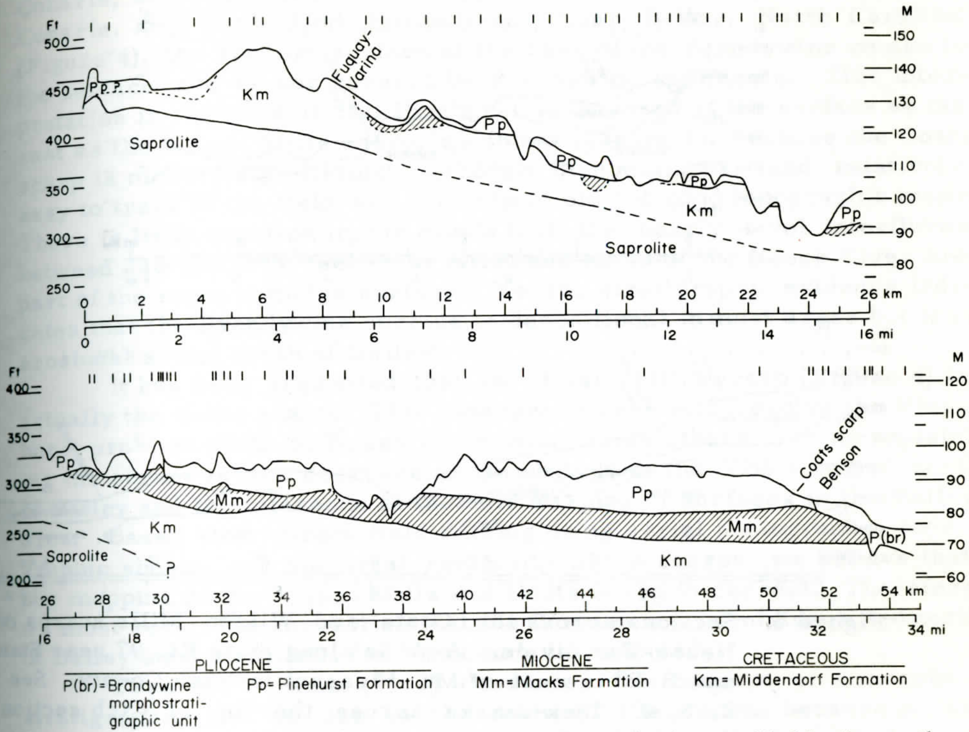

Figure 5. Section along the Neuse-Cape Fear drainage divide from the Piedmont border, about 5 miles northwest of Fuquay-Varina, southwest to Benson. See Figures 2, 3, or 4. Tick-marks across the top are drill hole locations.

Coats scarp. At the Wilson Mills scarp, the base of the sediments of the Coharie morphostratigraphic unit (msu) is inset considerably below the extremely rough base of the Pinehurst Formation (Figure 6). In most areas, the Coats scarp truncates the Pinehurst and Macks Formation (Daniels et al., 1966, Fig. 3), but near Bailey, it is the Wilson Mills scarp, not the Coats, that truncates these sediments (Figure 6). Immediately south of Bailey, drill hole data and surface mapping indicate that large areas of the Macks Formation are exposed on the Coats scarp and that the base of the Pinehurst is above the surface of the Brandywine sediments.

We recognize that the extension of the Pinehurst Formation east of the Coats scarp north of Bailey imposes difficulties in interpretation. But we were unable to fine any evidence of a break in sedimentation at and east of the Coats scarp. Until such a break is found, the sediments east and west of the Coats scarp in the area north of Bailey are one lithostratigraphic unit. It is possible that the Brandywine msu overlies the Pinehurst Formation east of the Coats scarp, but if so it was not detectable by the methods used.

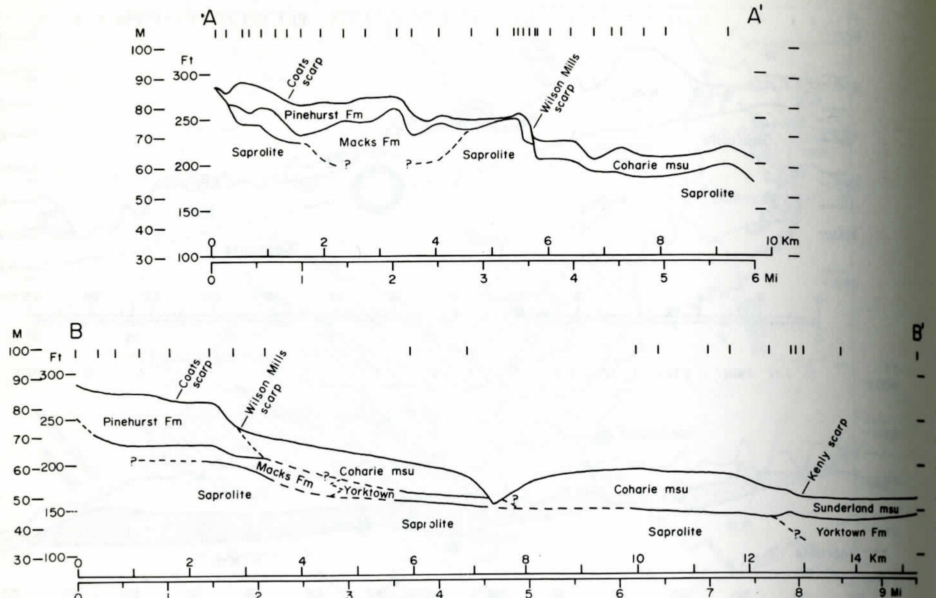

Figure 6. Sections across the Coats and Wilson Mills scarps on the Neuse-Tar divide. A-A' is along State Rt. 97 near Stanhope and B-B' is east of Mt. Pleasant in Nash County. See Fig. 2, 3, 4. Tick-marks across the top of each section are drill hole locations.

INTERPRETATIONS

From the information available in Figures 5 and 6 and elsewhere (Daniels et al., 1966), there can be little doubt that in most areas the Coats scarp marks the eastward limit of the Pinehurst, whereas near Bailey and Pinehurst extends eastward to the Wilson Mills scarp. Because the Pinehurst is a fluvial unit deposited on an erosion surface (ibid), this range in altitude is not surprising nor is the absence of Pinehurst from large areas north of the Neuse River between altitudes of 300 and 400 feet surprising. It is not necessary to deposit a fluvial unit either at the same altitudinal range throughout its distribution or to deposit sediment at all places within the range of altitude of the unit. If our interpretation of the Pinehurst is correct, then two different interpretations of the Brandywine msu must be made depending upon geographic location.

In most areas, the erosion surface at the base of the sediments of the Brandywine msu truncates the Pinehurst and Macks Formations and rises to the surface as the Coats scarp (Fig. 3, Daniels et al., 1966). In these areas, the sediments of the Brandywine msu south of the Neuse are part of an unnamed formation that includes the Brandywine

Coharie, and Sunderland msu, or parts of Stephenson's (1912) Lafayette, Coharie, and Sunderland Formations. Near Bailey, North Carolina (Figure 4), the erosion surface at the base of the Brandywine comes to the surface and is not covered by Brandywine sediments. This interpretation is required if the Pinehurst is exposed at the surface as far east as the Wilson Mills scarp, as shown (Figure 4), because the Coats scarp is moderately distinct, although discontinuous, and relatively easy to trace in the field with the help of old but good topographic maps. There is little question in our minds that the nearly level interfluves between 240 and 275 feet both north and south of the Neuse River are part of the same complex surface. Yet the stratigraphic evidence indicates that the Brandywine surface is depositional in most areas but it is erosional at and north of Bailey.

It has been suggested that the Wilson Mills scarp (Figure 4) is actually the Coats scarp. This interpretation would resolve the Pinehurst problem north of Bailey but it would raise others such as explaining the abrupt 10 foot decrease in toe altitude of the "Coats scarp" north of Bailey and the inference of tectonic warping of surfaces in the Bailey area. Based upon direct field tracing using aerial photographs, topographic sheets, and the areal continuity of the scarps, we believe that our mapping of the Wilson Mills and Coats scarp is correct. Therefore, we must reject the suggestion that the Coats scarp is misidentified north of Bailey until additional field evidence is collected.

Frye and Willman (1962) developed the concept of a morphostratigraphic unit to use in the midwest Pleistocene because strict stratigraphic nomenclature and concepts would not allow recognition of units important in the Pleistocene history of the area. An msu is recognized and mapped largely on its surface form, not on the distinctiveness of the underlying material. As such, an msu has a geomorphic bias that is not allowed in standard stratigraphy. But sedimentary bodies are the basis for definition of a morphostratigraphic unit and although erosion surfaces are not excluded they are not a primary consideration in the definition.

Because there is more area of depositional than erosional surface within the Brandywine msu, we feel that we can apply the term morphostratigraphic unit to the sediments underlying both erosional and depositional surfaces without violating the original concepts of the idea. By allowing this kind of range or flexibility in a morphostratigraphic unit, it is possible to have the Brandywine msu south of the Neuse as part of a yet unnamed formation that encompasses most of the middle Coastal Plain (Daniels et al., 1972), and the Brandywine msu north of Bailey be part of the fluvial member of the Pinehurst Formation. This violates neither the stratigraphic nor the geomorphic evidence presented. But using both a stratigraphic and geomorphic approach helps considerably in trying to understand an area.

SUMMARY

The upper Coastal Plain surficial sediments range from the Cretaceous Middendorf to the late Miocene or Pliocene fluvial member of the Pinehurst. Above 420 to 450 feet on the Neuse-Cape Fear Divide, the Middendorf is exposed, and it apparently never was buried by the younger Macks and Pinehurst Formations. The fluvial member of the Pinehurst Formation is the most extensive surficial deposit below 420 feet, although post-depositional erosion has removed it from several topographic lows along the divide. South of Bailey, the eastward limit of the Pinehurst Formation is the Coats scarp, but north of Bailey, the Pinehurst extends into the middle Coastal Plain under the erosional element of the Brandywine morphostratigraphic unit. The eastward limit of the Pinehurst Formation between Bailey and the Tar River is near the Wilson Mills scarp.

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LITHOSTRATIGRAPHY, DEPOSITIONAL ENVIRONMENTS, AND CONODONT BIOSTRATIGRAPHY OF THE MULDRAUGH FORMATION (MISSISSIPPIAN) IN SOUTHERN INDIANA AND NORTH-CENTRAL KENTUCKY

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ABSTRACT

The Muldraugh Formation is composed of three formal units presented in ascending order: the Floyds Knob Bed (new rank) consists of crinoid biosparites, oosparites, and biomicrites; the Edwardsville Member, composed of bedded and massive siltstones in the northeast, grades southwestward into interbedded shales and crinoidal carbonates; and the Ramp Creek Member, a complex of crinoidal carbonates and argillaceous dolomites commonly silicified. Two important glauconitic marker beds are in the Muldraugh Formation. The lower bed, previously known as the Floyds Knob glauconite, actually is at the base of the Edwardsville Member; the upper marker bed, recognized by Whitehead (1973), is at the base of the Ramp Creek Member.

The Muldraugh Formation was deposited in an outer platform - upper slope position near the southwestern limit of the Borden delta. Interformational clasts of early diagenetic siltstone concretions found in the Floyds Knob Bed indicate a period of post-Borden - pre-Muldraugh erosion. Crinoid biosparites and oosparites in the Floyds Knob Bed represent the initial transgressive phase of an important (eustatic?) sea level rise. Continued sea level rise trapped detrital sediment in near-shore areas to the northeast and permitted the development of a slowly accumulated, glauconite-strewn surface at the base of the Edwardsville Member. In Edwardsville time, silt and fine sand, introduced by minor turbidity flows (paleocurrent mean towards S30°W) and normal marine currents, were deposited in the northeast while shales were being deposited in the southwest. Associated with the shales were isolated crinoid colonies that grew in this area of low to moderate sedimentation. Edwardsville sedimentation was halted by another (eustatic?) sea level rise during which time another glauconite zone formed. In Ramp Creek time, crinoid meadows growing at the platform-slope break contributed

skeletal debris to carbonate muds accumulating in lagoonal environments on the platform and to carbonate muds being deposited in deeper waters seaward of the shelf break.

The conodont fauna of the Muldraugh Formation is dominated by Gnathodus texanus with a small, but significant, amount of Taphrognathus varians. This fauna indicates the Muldraugh Formation lies within the Gnathodus texanus - Taphrognathus Zone and correlates biostratigraphically with the Keokuk Limestone of the type Mississippian of the United States.

INTRODUCTION

The Muldraugh Formation in southern Indiana and north-central Kentucky is transitional between the underlying, terrigenous detrital Knobs Megagroup (Swann and Willman, 1961) and the overlying, carbonate Mammoth Cave Limestone Megagroup (Swann and Willman, 1961). The Muldraugh Formation is especially interesting because the unit may contain evidence of important eustatic changes and/or tectonic movements associated with sea floor spreading, which could account for the major change from clastic to carbonate deposition in the final stages of the Acadian orogeny.

This investigation is primarily a field study of the stratigraphic distribution and relationship of rock units to one another. Lithostratigraphic and sedimentologic conclusions are based on a study of 91 closely spaced outcrops (Figure 1). Detailed measured sections of these outcrops are presented in Whitehead (1976, Appendix B).

In processing 183 kilograms of rock from the Muldraugh Formation, 2007 conodonts were recovered and formed the basis for this study. Nicoll (1971) and Nicoll and Rexroad (1975) described the conodonts from the Ramp Creek Member of the Muldraugh Formation, but they did not investigate the Floyds Knob Bed. Combining the information of this report, a detailed study of the Floyds Knob fauna, and that of Nicoll and Rexroad (1975), a more confident biostratigraphic correlation with the type Mississippian can be made.

Acknowledgments

I wish to extend my thanks to: Charles Collinson, who directed my M. S. thesis upon which this paper is based; Roy C. Kepferle and Jerry A. Lineback for helpful discussions on Mississippian stratigraphy; Mr. and Mrs. Neil H. Whitehead, Jr. for financial support of my fieldwork; and Mrs. Ann Waybright Whitehead, who did editing, typing, and drafting for this report. John M. Dennison and Roy C. Kepferle critically read this report in manuscript form and offered many helpful suggestions for improvement.

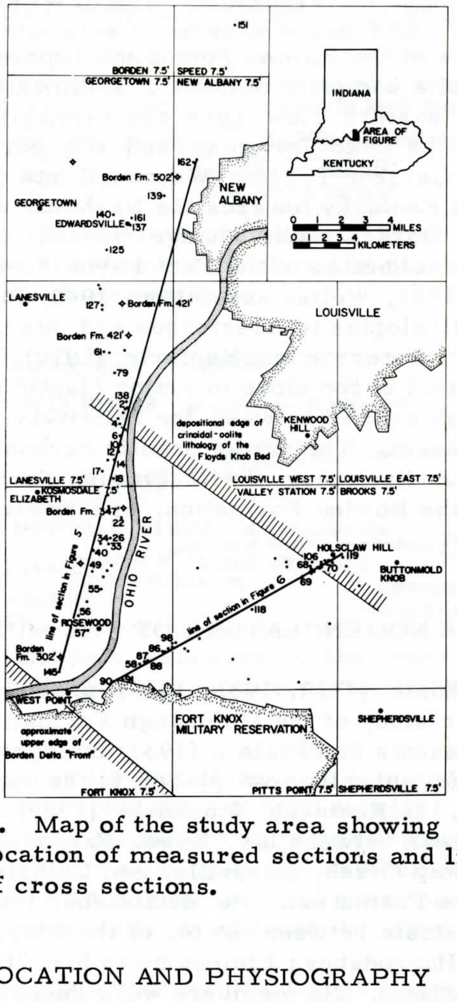

Figure 1. Map of the study area showing the location of measured sections and lines of cross sections.

LOCATION AND PHYSIOGRAPHY

The area under study (Figure 1) in Indiana comprises parts of Clark, Floyd, and Harrison Counties, and in Kentucky, parts of Jefferson, Bullitt, and Hardin Counties. This area, which lies on the west flank of the Cincinnati arch, is underlain by rock units that are flat-lying, with dips usually between 20 to 60 ft/mi (4 to 11 m/km). Borden and Muldraugh rocks are exposed along a steep, east-facing escarpment called the Knobstone Escarpment in Indiana and the Muldraugh Escarpment in Kentucky. Along the escarpment front, the relief ranges from 300 to 400 feet (91 to 122 m), and the topography is rugged. Thousands of steep-gradient ravines, which Stockdale (1931) aptly calls flume ravines, groove the hillsides, offering excellent exposures.

STRATIGRAPHIC SETTING

Rocks of the Borden Formation represent deposition along the distal edge of a complex of deltaic sediments that began in the Middle Devonian in eastern New York and Pennsylvania and prograded westward during the Lake Devonian and the early Mississippian Kinderhookian and Osagean Epochs. Westward into the Illinois basin and southward across Kentucky towards the Nashville dome area, the fine-grained, detrital Borden rocks change through a narrow zone into cherty, argillaceous dolomites of the Fort Payne Formation. Previous workers (Stockdale, 1939, Weller and Sutton, 1940; Pinsak, 1957) believed these two major lithologies to intertongue and intergrade; however, Lineback (1966, 1969), Peterson and Kepferle (1970), and Kepferle (1972c) demonstrated that at the close of major clastic deposition, an abrupt topographic break existed between the relatively shallow shelf and the relatively deep basin. The predominantly carbonate rocks that filled this starved basin (including the Muldraugh Formation), although laterally adjacent to the Borden Formation, are not contemporaneous with the Borden.

PREVIOUS NOMENCLATURE OF THE MULDRAUGH FORMATION

Stockdale (1929, 1931, 1939) provides the only regional lithostratigraphic study of the Muldraugh Formation in Kentucky and Indiana. Figure 2 presents Stockdale's (1931) interpretation of stratigraphic relationships of subunits now placed in the upper Bordan and Muldraugh Formations. In Kentucky, Stockdale (1939) was unable to recognize a lithologic break between his "Lower Harrodsburg Limestone" (composed of the Ramp Creek, Leesville, and Guthrie Creek Members) and the Edwardsville Formation. He established the Muldraugh Formation to include the strata between the top of the Floyds Knob Formation and the base of the Harrodsburg Limestone (= his "Upper Harrodsburg Limestone" of Indiana). No members were recognized within the Muldraugh Formation; however, the entire unit cropping out between Jefferson and Larue Counties, Kentucky, was designated as the West Point Facies. Stockdale (1939, p. 203) extended the name West Point Facies of the Muldraugh Formation, into southern Harrison County, Indiana (as far north as the middle of the Kosmosdale 7. 5-minute quadrangle), in which he included the following units: Stewarts Landing Facies of the Edwardsville Formation and the Ramp Creek, Leesville, and Guthrie Creek Members of his "Lower Harrodsburg Limestone".

Smith (1965) considered the interval between the base of the Floyds Knob Formation of Stockdale (1931) and the top of the Ramp Creek Member of the "Lower Harrodsburg Limestone" (Stockdale, 1931) as the Muldraugh Formation and extended the use of this term throughout the surface outcrop in Indiana. In addition, he reduced the rank of

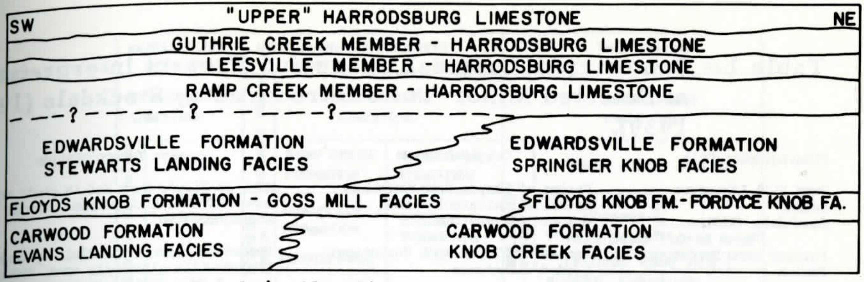

Figure 2. Stockdale's (1931) interpretation of stratigraphic relationships of subunits now placed in the upper Borden and Muldraugh Formations.

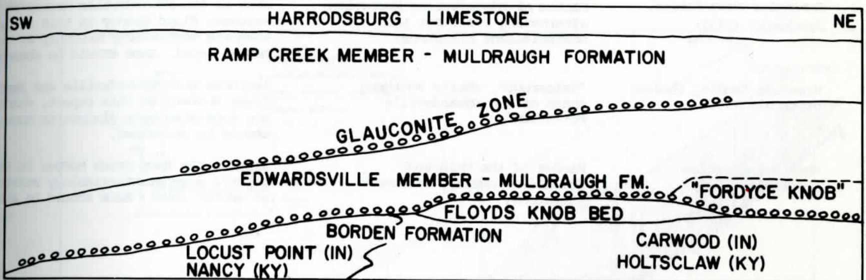

Figure 3. The stratigraphic relations of the upper Border and Muldraugh Formations as interpreted by Whitehead.

the Floyds Knob and Edwardsville from formation to member.

The basic lithostratigraphic units of Stockdale (1929, 1931, 1939) remain in use today; however, his numerous facies names only confuse an already adequate terminology existing at the member and formational levels. Table 1 summarizes briefly the original usage and present interpretation of selected minor units introduced by Stockdale (1931, 1939).

The nomenclatural history of the Muldraugh Formation is quite complex, and the reader is referred to summaries and discussions found in Smith (1965), Lineback (1966), Shaver and others (1970), Kepferle (1971), and Nicoll and Rexroad (1975). See Figure 4 for a comparison between the stratigraphic nomenclature of Kepferle (1971), Nicoll and Rexroad (1975), and this report.

LITHOSTRATIGRAPHY OF THE MULDRAUGH FORMATION

The Muldraugh Formation rests disconformably on the Borden Formation and is succeeded conformably by the Harrodsburg Limestone. In Indiana, the Muldraugh Formation ranges from 75 to 100 ft (23 to

Table 1. Summary of Original Usage and Present Interpretation of Selected Minor Units Introduced by Stockdale (1931, 1939).

Straitgraphic Unit	Original Usage	Present Interpretation
Goss Mill Limestone Facies Stockdale (1931)	Facies of Floyds Knob Formation; crinoidal calcarenite and dolomitic limestone.	Same as Floyds Knob Bed in study area; name has no useful prupose, should be abandoned.
Fordyce Knob Sandstone Facies Stockdale (1931)	Facies of Floyds Knob Formation; bedded siltstones.	Bedded siltstones underlain by basal Edwardsville glauconite zone, therefore part of Edwardsville Member; name should be abandoned.
Evans Landing Facies Stockdale (1931)	Facies of Carwood Formation; shale, sandy, slightly calcareous in upper part.	At type section, upper part is Edwardsville, lower part is Locust Point lithology; name should be abandoned.
Springler Knob Facies Stockdale (1931)	Facies of Edwardsville Formation; alternating resistant and nonresisteant siltstone.	Same as the Edwardsville Member in southern Floyd County in this study; adequate terminology existing at the member level; name should be abandoned.
Stewarts Landing Facies Stockdale (1931)	"calcareous, cherty southern phase of the Edwardsville Formation".	Includes both Edwardsville and Ramp Creek Members of this report, which are separated by a glauconite zone; name should be abandoned.
West Point Facies Stockdale (1939)	Facies of the Muldraugh Formation; cherty, siliceous limestone.	Same as the Ramp Creek Member in this report; adequate terminology existing at member level; name should be abandoned.

30 m) thick; in Kentucky, based on less detailed study, the Muldraugh Formation ranges from 55 ft (17 m) thick in the northeast to over 100 ft (30 m) in the southwest. Most of the thickening occurs in the southwesternmost part of the study area and is associated with the abrupt thinning of the underlying Borden Formation. Figure 3 shows stratigraphic relations of the Muldraugh Formation as interpreted from this study. Figure 5 (Indiana) and Figure 6 (Kentucky) are lithologic cross sections of the Muldraugh Formation.

Floyds Knob Bed

This unit is a fossiliferous carbonate composed of two principle lithologies. In Indiana, the mean thickness is about 2 ft. (0.6 m). In Kentucky, the mean thickness is about 2.5 ft. (0.8 m).

The lower subunit is an oosparite to crinoid biosparite. This subunit is predominantly horizontally laminated, although there is small scale cross-bedding in outcrops in the Brooks and Valley Station 7.5-minute quadrangles. Scattered grains of pelletoidal glauconite are present along with some minor glauconite replacement of echinoderm skeletal material. Calcareous siltstone, oosparite, and biosparite clasts (ranging up to small cobble size - 150 mm) are common at many localities. The basal contact of this subunit is sharp and irregular at all localities studied, with no glauconite zone noted beneath this subunit. The upper contact with the overlying, biomicritic subunit of the Floyds Knob Bed is, in most places, sharp and seemingly conformable. The mean thickness of the oosparite to crinoid biosparite subunit in Indiana

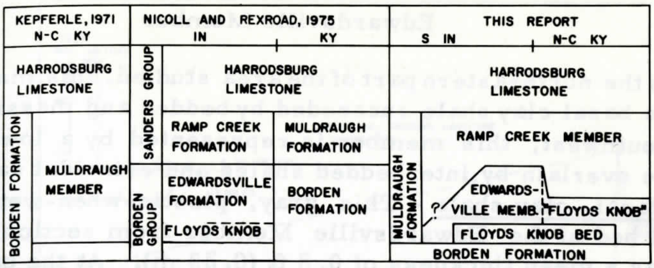

Figure 4. Comparison of stratigraphic nomenclature of Kepferle (1971) and Nicoll and Rexroad (1975) with that presented in this report.

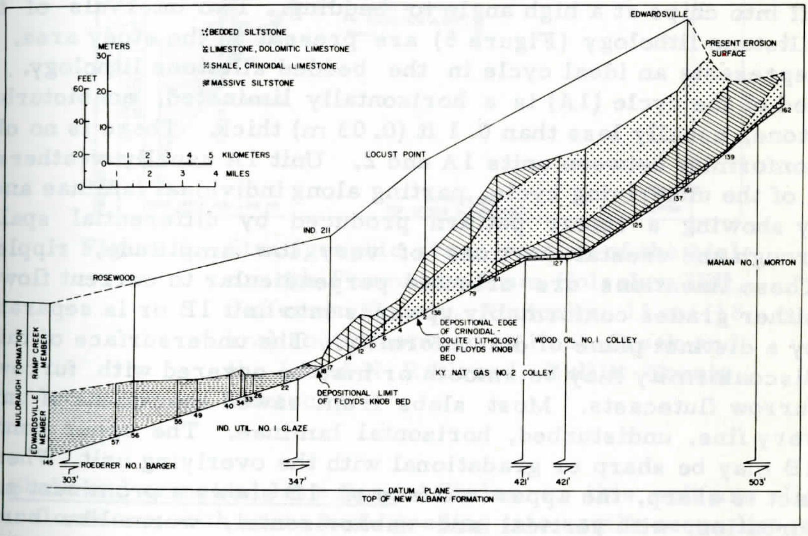

Figure 5. Stratigraphic cross section of the Muldraugh Formation from Floyds Knob, Floyd County, Indiana, 23 mi (37 km) southwestward to Evans Landing, Harrison County, Indiana.

is one foot (0.3 m). In Kentucky, this subunit is the only representative of the Floyds Knob Bed.

The upper subunit of the Floyds Knob Bed is a slightly dolomitic biomicrite, with a distinctive reddish-brown color on weathered surfaces. Fossils are common to abundant in this subunit; especially characteristic are large, thick-shelled brachiopods and phosphatic steinkerns of high-spired gastropods. The unit is overlain disconformably by a glauconite zone in the basal part of the Edwardsville Member. The mean thickness of this subunit is about 1.5 ft (0.5 m).

Edwardsville Member

In the northeastern part of the area studied, this member is composed of a basal clay shale succeeded by bedded and massive stilstones. To the southwest, this member is represented by a lower, silicified limestone overlain by interbedded shales and crinoidal carbonates.

Basal, clay shale - This gray, plastic-when-wet shale forms the basal bed of the Edwardsville Member from section 151 to section 18 and has a mean thickness of 0.8 ft (0.23 m). At the base of (and included within) this subunit is a zone of greenish-black, pelletoidal glauconite.

Bedded siltstone - This lithology is composed of interbedded, ledge-forming siltstones and less resistant, shaly siltstones that spall off into chips at a high angle to bedding. Two intervals of the bedded siltstone lithology (Figure 5) are present in the study area. Figure 7 represents an ideal cycle in the bedded siltstone lithology. The basal bed of the cycle (1A) is a horizontally liminated, nonbioturbated siltstone, usually less than 0.1 ft (0.03 m) thick. There is no obvious disconformity between units 1A and 2. Unit 1A usually weathers with unit 2 of the underlying cycle, parting along individual laminae and frequently showing a linear pattern produced by differential spalling of the trough and crestal portions of very low amplitude, ripple laminae. These lineations are oriented perpendicular to current flow. Unit 1A either grades conformably upwards into unit 1B or is separated from it by a distinct plane of disconformity. The undersurface of this plane of disconformity may be smooth or may be covered with furrowcasts and furrow flutecasts. Most slabs from sawed specimens of unit 1B show very fine, undisturbed, horizontal laminae. The upper contact of unit 1B may be sharp or gradational with the overlying unit. Where the contact is sharp, the upper surface of 1B shows a prominent zone of bioturbation, with vertical and subhorizontal, worm-like burrows and Zoophycos as the dominant trace fossils. Unit 2, a shaly siltstone with rare laminations or bedding structures, contains abundant, trace fossil bioturbation structures.

Massive siltstone - This siltstone, upon weathering, yields angular, thin slabs that spall off at a high angle to bedding. Trace fossil abundance demonstrates intensive bioturbation in this unit. The trace fo‿sil population is dominated by small fodinichnia (see Stockdale, 1939, Plate 4, Figure 1); also present are larger, vertical and subhorizontal, cylindrical burrows and Zoophycos.

Siliceous dolomicrite - This lithology occurs at the base of the Edwardsville Member in the southwestern part of the study area in Indiana and Kentucky. A glauconite zone at the base of this subunit is correlated lithostratigraphically with the glauconite zone on top of the Floyds Knob Bed to the northeast. This lithology is somewhat erratic in occurrence and is included within the shale and crinoidal limestone subunit on Figures 5 and 6.

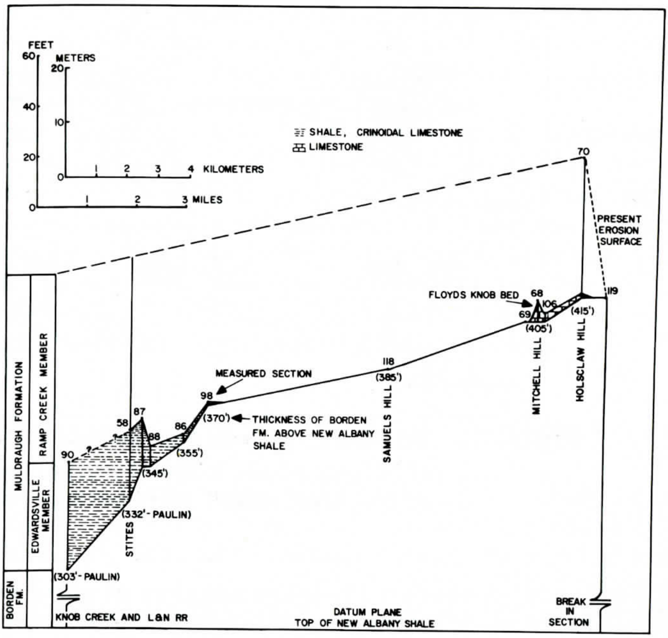

Figure 6. Stratigraphic cross section of the Muldraugh Formation from Holsclaw Hill, Jefferson County, Kentucky, 11 mi (18 km) southwestward to Knob Creek and the L and N Railroad, Bullitt County, Kentucky.

Shale and crinoidal limestone - Shales in this unit are slightly silty to calcareous, with some beds grading into argillaceous limestone. The limestone is a crinoid, bryozoan biosparite. The upper and lower contacts of the limestone beds are sharp with bedding thickness varying considerably between some closely-spaced outcrops. At several sections, limestone beds are closely interbedded with shales, suggesting probable lateral gradation of the two rock types. No primary bedding structures were noted in the limestones; however, at several sections, beds showed a slight amount of initial dip, produced by sedimentation on a minor depositional slope.

Ramp Creek Member

The dominant lithology of this member is argillaceous dolomicrite which grades to a calcitic, silty dolomite. Geodes are common to abundant and contain interior fillings of quartz, dolomite, and gypsum. Gypsum-filled geodes seem to be associated with less-weathered exposures. At some outcrops chertification is extensive and replaces

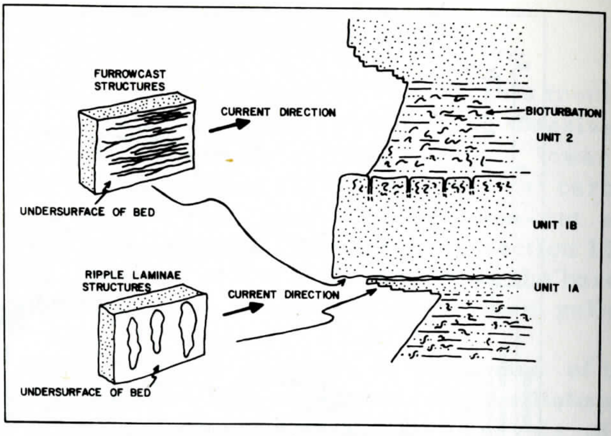

Figure 7. An ideal cycle in the bedded siltstone lithology of the Edwardsville Member of the Muldraugh Formation.

entire beds, and at section 17, large masses of solid chert grade laterally into calcitic, silty dolomite with abundant chert and common geodes. Much of the dolomicrite shows bioturbation, as indicated by common fodinichnia.

Crinoid biosparites are complexly interstratified with the argillaceous dolomicrite lithology. Abrupt lateral variation between the two above lithologies occurs in sections 140 and 161; these lithologies also grade vertically into each other. At section 17, five cyclic alternations of crinoid biosparite with argillaceous dolomicrite are well-developed in the upper Ramp Creek Member. Here the crinoid biosparite has a sharp, disconformable, basal contact. Horizontal, feeding trackways, filled with crinoid biosparite, are developed on the surface of the underlying, argillaceous dolomicrite. The upper contact of these crinoid biosparites is gradational.

The lower contact of the Ramp Creek Member is marked by a well-developed, glauconite zone. Abundant, greenish-black, pelletoidal glauconite, as a nearly pure seam or in a matrix of dolomicrite, constitutes the zone. In Indiana, this glauconite zone is well-developed southward from section 16 to section 56 (approximately 11 mi (18 km)). Northeast of section 61, the basal Ramp Creek glauconite zone is much less developed; however, the sharp change up the section from detrital to carbonate allows recognition of the Ramp Creek contact. Southwest of section 56, the glauconite zone is apparently missing. In this area, near the Border delta front, the Edwardsville Member and the Ramp Creek Member are lithologically similar, so separating them is difficult. In Kentucky, a similar situation occurs southwest of section 58 where the upper glauconite zone, which correlates lithostratigraphically with the basal Ramp Creek glauconite, is absent. In Kentucky, between sections 58 and 98, the basal Ramp Creek glauconite zone is present. Northeast of section 98, intervening strata of the Edwardsville Member are absent, and the basal Edwardsville glauconite zone and the basal Ramp Creek glauconite zone apparently converge.

The upper part of the Muldraugh Formation comprises a series of alternating, argillaceous dolomites and crinoid biosparites, which grade into each other vertically and laterally. In this report, the Leesville and Guthrie Creek units (Stockdale, 1929) are not recognized; possible equivalents are included in the Ramp Creek Member. The upper contact of the Ramp Creek Member (and of the Muldraugh Formation) is placed at the base of the first persistent, crinoid biosparite above which there are no major argillaceous dolomite beds.

PRESENT NOMENCLATURE OF THE MULDRAUGH FORMATION

The Muldraugh Formation as defined in this study (Figure 3) includes three formal subunits; they are, in ascending order: the Floyds Knob Bed (new rank), the Edwardsville Member, and the Ramp Creek Member.

In Kentucky, there is a sharp lithologic contrast between the terrigenous detrital Borden rocks and the carbonate-rich Muldraugh section. In Indiana, the lithologic differences are less pronounced, yet the important pre-Floyds Knob Bed disconformity separates the two units. For these reasons, the Muldraugh Formation is separated from the Borden Formation in this report.

The Floyds Knob Bed is too thin to show as more than a line on a 1:24,000 scale map. For this reason it seems more appropriate to apply the rank of bed rather than member to this unit.

As a result of this study, in particular from the recognition of a second glauconite zone within the Muldraugh Formation of southern Indiana, it is evident that usage of the term "Floyds Knob" in central Kentucky is different from that in the type area of southern Indiana. As defined herein, the Floyds Knob Bed lies below any zone of pelletoidal glauconite and is restricted at present in Kentucky to the Brooks and Valley Station 7.5-minute quadrangles. Stockdale (1939) used the name "Floyds Knob" for a glauconitic bed or beds and the associated rock sequence. Kepferle (1971, 1972c) and Sedimentation Seminar (1972) apparently consider the "Floyds Knob" unit as bounded by an upper and lower glauconitic interval. Strata in part of the Kosmodale 7.5-minute quadrangle apparently link lithostratigraphically the Edwardsville Member of southern Indiana with the "Floyds Knob" of central Kentucky. Thus, the "Floyds Knob" as used by Stockdale (1939) and Kepferle (1971), includes the Floyds Knob Bed of this report as well as strata which are probably chronostratigraphically equivalent to the Edwardsville Member in its type area. On account of lithologic dissimilarities between the type Edwardsville and the "Floyds Knob" of Kentucky, it is not recommended that the term "Edwardsville" be used in a lithostratigraphic sense in Kentucky.

CONODONT BIOSTRATIGRAPHY

The presence in the Muldraugh Formation of the important platform species Gnathodus texanus Roundy and Taphrognathus varians Branson and Mehl indicates this rock unit lies within the Gnathodus texanus - Taphrognathus Zone (Collinson, Scott and Rexroad, 1962) or lower part of the Taphrognathus varians-Apatognathus Zone (Collinson, Scott and Rexroad, 1962). Collinson and others (1971) note the Gnathodus texanus - Taphrognathus Zone is confined essentially to the Keokuk Formation of the type Mississippian, although it may range locally downward into the upper part of the Burlington Limestone. The base of this zone is defined as the earliest abundant occurrence of G. texanus; the top of this zone is defined as the earliest abundant occurrence of T. varians.

Collinson and others (1971) modified the lower limit of the Taphrognathus varians - Apatognathus Zone to include strata marked by the earliest common occurrence of T. varians. This zone is found in the Warsaw, Salem, and lower St. Louis rock units of the Mississippi Valley. Of especial interest to this study is the common occurrence of G. texanus in the lower part of this zone.

Rexroad and Collinson (1965), in their study of the Keokuk, Warsaw, and Salem Formations in Iowa and Illinois, show an upwards increase in percentage of T. varians and a decrease in the percentage of G. texanus. Their data are summarized in Table 2. Similar data were compiled for the Indiana and Kentucky outcrops (Table 2); the data for the Floyds Knob Bed and the Edwardsville Member were obtained from Whitehead (1976, Tables 4-7); data for the other units were obtained from Nicoll (1971, Tables 3, 8, 11-14, and 17). The Ramp Creek Member data (Table 2) includes information from strata recognized by some workers as the Leesville and Guthrie Creek Members.

Table 2 shows that a similar relationship exists between the Salem, Warsaw, and Koekuk of the type Mississippian and the Salem, Harrodsburg, and Muldraugh Formations of Indiana and Kentucky. For this reason, I correlate in a biostratigraphic sense the Keokuk Limestone with the Muldraugh Formation. Nicoll and Rexroad (1975) conclude that the base of the Taphrognathus varians - Apatognathus Zone lies in the upper part of the Ramp Creek Member; however, abundance data, from Nicoll (1971) and summarized in Table 2, indicate the first abundant occurrence of T. varians is in the Harrodsburg Formation.

Some evidence is present in southern Indiana and north-central Kentucky to suggest the abundance of T. varians is related to the presence of a crinoid biosparite and/or oosparite (indicating shallow, agitated water). The Floyds Knob Bed, an oosparite to crinoid biosparite,

Table 2. Percentage of Gnathodus texanus and Taphrognathus varians of the Total Recovered Conodont Population.

SE Iowa and W Illinois	T. v.	G. t.	S. Indiana and N-C Kentucky	T. v.	G. t.
T. varians - Apatognathus Zone					
Salem Fm.	58%	3%	Salem Fm.	66%	0%
Warsaw Fm.	20%	44%	Harrodsburg Ls.	32%	28%
G. texanus - Taphrognathus Zone					
Keokuk Limestone	7%	78%	Ramp Creek Member	2%	82%
			Edwardsville Member	1%	72%
			Floyds Knob Bed	11%	74%

contains 11 percent T. varians. The overlying, basal Edwardsville glauconite zone, here interpreted as a deposit of increasing water depth, contains less than one percent of T. varians. The remaining part of the Edwardsville Member has no reported T. varians. The basal Ramp Creek glauconite, another increasing-depth deposit, has no reported T. varians. The rest of the Ramp Creek Member is mostly an argillaceous, cherty dolomicrite and has two percent T. varians. Beginning with the Harrodsburg Limestone, a crinoid biosparite, and continuing into the Salem Limestone, a foraminiferal to crinoid biosparite, the T. varians percentage increases abruptly from 32 to 66 percent. Finally, in the St. Louis Limestone, a dark micrite, T. varians decreases to extinction.

A summary of conodont distribution in the Muldraugh Formation (Whitehead, 1976) is presented in Table 3. These data should be combined with the information presented by Nicoll and Rexroad (1975) to furnish a more complete picture of the Muldraugh fauna. The only previous mention of conodonts from the Floyds Knob Bed was by Gates and Rexroad (1969, p. 22), who reported only that the conodonts were of Keokuk age. Conodonts reported from the Edwardsville Member in Table 3 are mainly from the glauconite zone at the base of that unit. Conodonts from the Ramp Creek Member in Table 3 are from the glauconite zone at the base of this unit in the area where the Edwardsville Member is missing by nondeposition. The fauna listed in Table 3 are illustrated and described systematically in Whitehead (1976).

Table 3. Summary of the Recovery and Distribution of Conodonts from the Muldraugh Formation (data from Whitehead (1976, Tables 2, 4-7)). This table was compiled from complete conodont elements and selected, unique conodont fragments (to prevent fragmentary conodont elements from being counted twice).
* indicates a downward extension in range from that previously reported by Nicoll and Rexroad (1975)
# indicates this taxon was not recorded by Nicoll and Rexroad (1975)
x indicates taxon was present but not counted
@ indicates value not representative of average conodonts/kg

	Floyds Knob Bed	Edwardsville Member	Ramp Creek Member
Gnathodus texanus	801	388	306
G. sp. cf. G. cuneiformis			19#
G. sp. A		x#	
Hibbardella colobus	3#	5#	
Hindeodella sp.	x	x	x
Lingonidina sp.	26	33	10
L. roundyi	18*	33*	7
Lonchodina sp.	19	17	2
L. paraclarki	17*	42*	6
L. paraclaviger	11*	7*	31
Neoprinoiodus acamplyus	1#		
N. loxus	4#		
N. tulensis	2*		
N. sp. cf. N. conjunctus	3		1
Ozarkodina compressa	17*		
O. roundyi	13*	11	3
Spathognathodus coalescens	22#		
S. sp. A			7#
Synprioniodina sp.		1	
Taphrognathus varians	117*	4*	
Total conodonts recovered	1074	541	392
Weight of sample (kg)	105.5	61.9	15.6
Conodonts/kg	10	9@	25@

DEPOSITIONAL HISTORY OF THE MULDRAUGH FORMATION

Pre-Muldraugh Depositional Topography

An understanding of the depositional topography that existed after cessation of Borden sedimentation is necessary to provide the setting of the depositional environments in the overlying Muldraugh Formation. Peterson and Kepferle (1970) and Kepferle (1972c) report the presence of major depositional topography at the close of Borden deposition in

Kentucky and Indiana as determined from variations in the stratigraphic interval between the lower contact of the Muldraugh Formation and the base of the underlying Borden Formation. The basal Muldraugh is marked by a burrowed, glauconite pellet-strewn surface that represents an interval of slight or no deposition. From this preserved sea floor surface, the approximate topography of the Borden delta front may be reconstructed using an isopach map or cross section of the Borden Formation (Figures 5 and 6). Thus, at the end of Borden deposition in southern Indiana and north-central Kentucky, the delta stood a minimum of 500 ft (152 m) (and, allowing for compaction, as much as 600 ft (183 m) above the basin floor. Water depths above the delta, although virtually unknown, may have been as much as 50 to 200 ft (15 to 61 m). An open sea stretched southwestward, with water depths of perhaps 650 to 800 ft (198 to 244 m), opening into the Ouachita geosyncline. In the study area, the Muldraugh Formation was deposited on the upper slope and outer platform of the delta. The upper part of the slope dips to the southwest at 22 ft/mi (4m/km), while the outer platform has slopes ranging from 6. 5 to 8. 5 ft/mi (1. 2 to 1. 6 m/km). All of these water depth and slope figures are based on the simplistic assumption that the pre-Borden sea floor was essentially level.

Post-Borden Pre-Muldraugh Erosion

Preserved in the Floyds Knob Bed are clasts of calcareous siltstone (Figure 8). The diameters of the clasts (using the Wentworth scale) range from small pebbles to large cobbles (150 mm, maximum diameter). The clasts are predominantly spherical to subspherical; however, a significant number of clasts show typical concretionary forms; these include elliptical forms with irregular, lobed projects (amoebiform and dumbell- or barbell-shaped forms. Most clasts contain fossils such as brachiopods, fenestrate bryozoans, and crinoid columnals; however, the fossils are incidental to the clasts and are not of the nucleus type commonly reported in concretions. Calcareous siltstone clasts in the Floyds Knob Bed have been reported from a number of localities in southern Indiana and north-central Kentucky: Jackson County, Indiana (Suttner and Hattin, 1973, p. 105), Clark and Floyd Counties, Indiana (Stockdale, 1931), and Jefferson County, Kentucky (Stockdale, 1939).

The Floyds Knob clasts are very similar to Quaternary occurrences of reworked early diagenetic concretions. Rusnak (1960, Figure 8) shows sandstone concretion gravels derived from erosion of unconsolidated Pleistocene outcrops along the mainland shoreline of Laguna Madre, Texas. Johnston (1921) and Garrison and others (1969) report similar concretions found at or near the sediment-water interface on the floors of distributary channels from the Holocene Fraser River Delta, British Columbia.

The following evidence supports a reworked early diagenetic

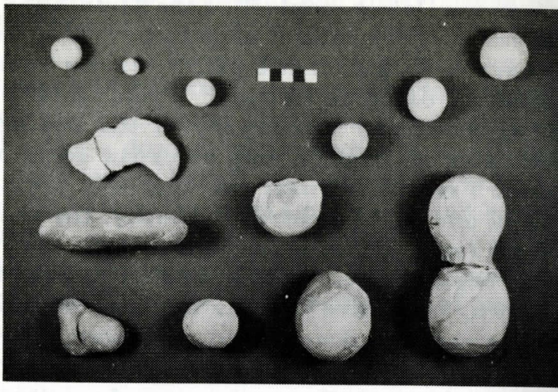

Figure 8. Interformational clasts from the Floyds Knob Bed of the Muldraugh Formation. Bar is 5 cm.

concretion origin for the calcareous siltstone clasts in the Floyds Knob Bed: (1) the contact of the Floyds Knob Bed with the underlying unit is erosional; (2) in Indiana, in the Carwood Formation (Stockdale, 1931, p. 156), and in Kentucky, in the Holtsclaw Member (Kepferle, 1972a, b), calcareous concretions are locally common; (3) there is a close similarity between the lithologic composition and extermal morphology of the clasts in the Floyds Knob Bed and the calcareous siltstone concretions in the Carwood Formation; (4) there is an absence of noncalcareous siltstone fragments in the Floyds Knob Bed which resemble the Carwood or Holtsclaw lithology - which supports an unlithified state for these units while they were being eroded.

In summation, these calcareous siltstone clasts are believed to be derived from early diagenetic concretions formed a few feet below the sediment-water interface during deposition of the upper Borden Formation. After the end of Borden deposition, a period of erosion ensued, which removed an unknown amount of unlithified silt and clayey silts of the Borden Formation (at least the upper few feet, and possibly, as much as several tens of feet). The pebble- and cobble-sized concretions remained as a lag conglomerate and were incorporated into the overlying Floyds Knob Bed. The near absence of encrusted or bored concretions may indicate a short time interval during which these nodules were exposed on the sea floor before incorporation into the Floyds Knob Bed.

Depositional Environment of the Floyds Knob Bed

The lower oosparite to crinoid biosparite subunit of the Floyds Knob Bed contains many features indicating a warm, shallow water, agitated, high energy environment, including cleanly washed skeletal

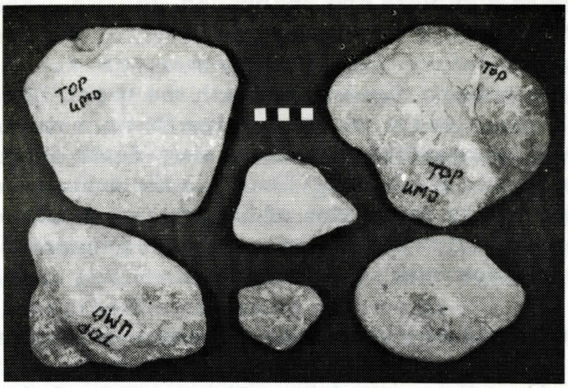

Figure 9. Intraformational clasts from the Floyds Knob Bed of the Muldraugh Formation. Bar is 5 cm.

sands composed of abraded and fragmental, echinoderm, bryozoan, brachiopod, and mollusk debris. In addition, the presence of oolites are a criterion of intertidal and shallow subtidal, tropical waters, supersaturated with respect to calcium. Newell and others (1960), from studies on the Bahama Bank, report oolites as having an optimum growth depth of 6 ft (2 m) or less. Bimodal cross-bedding, a characteristic feature of tidally dominated environments, was noted by Kepferle (1972c).

Intraformational clasts (Figure 9) present in the Floyds Knob Bed are composed of oosparite or biosparite. The size of the clasts ranges from small pebbles to small boulders (270 mm, maximum length), with an estimated mean size within the large pebble - small cobble range. The most common pebble shape is tabular to irregularly discoidal. Rare examples of subpherical clasts are also found. Clasts are well-rounded to subrounded, and some show a surface polish. At several outcrops, well-developed imbrication of the clasts is present. These limestone clasts are probably derived from disruptions of early diagenetically lithified sediment. Lithification may have been subaerial, as in the case of beachrock, or submarine. In any event, these clasts were formed at, or near, sea level, with storm waves probably producing the disrupting force.

The apparent southwestward depositional limit of the oosparite to crinoid biosparite subunit of the Floyds Knob Bed (Figures 1 and 10) strikes NW-SE. This trend is parallel to the strike of the Borden delta front, which lies some ten miles (16 km) to the southwest. The oosparite to biosparite lithology of the Floyds Knob Bed represents the initial phase of a transgressive deposit. Overlying the oosparite to biosparite subunit of the Floyds Knob Bed is a quiet water depositional limestone, as shown by its clayey and micritic composition and the occurrence of complete and unabraded fossils. Shallow water depositional features, such as algal stromatolites, birdseye structures, mud cracks, and intraformational conglomerates are absent. The occurrence of this subunit directly beneath the basal Edwardsville glauconite zone, which is clearly transgressive (but not necessarily significantly diachronous), strengthens an increasing-depth origin for the upper subunit of the Floyds Knob Bed. Thus, as the transgression continued, the water depth during deposition of the upper subunit of the Floyds Knob Bed increased, allowing calcareous muds generated elsewhere in a higher energy environment to settle out.

Depositional Environments of Recent and Ancient Glauconites

Perhaps the most interesting aspect of the Muldraugh Formation is the occurrence of one to several thin zones of abundant pelletoidal glauconite. Recent glauconites, forming on parts of the continental shelves today, permit an accurate interpretation of lithologically identical, ancient, glauconite deposits. It has been long recognized that glauconite forms in areas of slow or no sedimentation (Murray and Renard, 1891; Goldman, 1922). From modern studies of glauconite within Holocene sediments (Allen, 1965; Curry, 1960; Koldewijn, 1958; Nota, 1958; Van Andel, 1964, 1967; Van Andel and Veevers, 1967), a remarkably consistent pattern is evident. Glauconite in Holocene deposits occurs in areas of relict sediments and represents a deeper water addition to sands originally deposited as basal, transgressive, littoral sands following the Flandrian transgression. Fisher (1964), in a study of Eocene cyclic deposits of the northern Gulf Coast region, noted that although glauconites are quantitatively insignificant, they are important in regional correlation and in cycle analysis. The Eocene glauconites commonly occur at disconformities developed beneath minor to major marine transgressive units.

The physical and biological characteristics of the Muldraugh Formation glauconite zones are: (1) concentrations of dark green, sand-sized, pelletoidal glauconite within stratigraphically restricted intervals; (2) a sharp, unconformable, basal contact; (3) commonly an interval of extensive bioturbation beneath the glauconite zone dominated by channel-like trailways developed on the surface of unconformity; tubular vertical burrows; and <u>Zoophycos</u>; (4) concentrations of phosphate as nodules within the glauconite zone and as replacement crusts on underlying crinoidal and dolomitic limestones; (5) common occurrence of fish and linguloid brachiopod fragments; (6) abundant occurrence of conodonts (a probable indicator of slow or non deposition); and (7) upper contact of glauconite zones gradational and conformable.

Kepferle (1972c) recognized that the glauconites in the Muldraugh Formation represent a significant unconformity and postulates that this break indicates the destructive phase in deltaic deposition. Scruton (1960) explains that the destructive phase in deltaic sedimentation occurs when the major depositional area of the delta shifts. Compaction of sediments and subsidence continues in former depositional centers, with a resulting marine transgression over the sediments.

Stockdale (1939) traced one to several zones of glauconite in the position of the Muldraugh Formation from north-central Kentucky around the Jessamine dome into northeastern Kentucky. Abundant pelletoidal glauconite in a restricted stratigraphic interval occurs also in the upper Osagean rocks of Virginia (Bartlett, 1974) and Tennessee (Conant and Swanson, 1961; Hasson, 1972). The characteristics of these upper Osagean glauconites indicate a close similarity to Holocene glauconites in transgressive deposits on continental shelves and Cenozoic marine transgressive glauconites. The geographically widespread occurrence of this probably contemporaneous glauconitic interval in the upper Osagean suggests the Muldraugh Formation glauconites record a period of slow sedimentation following a major (eustatic?) sea level rise.

Depositional Environments in the Edwardsville Member

The glauconite zone at the base of the Edwardsville Member is interpreted as a transgressive (but not significantly diachronous) deposit. The rest of the Edwardsville Member is clearly regressive and represents the last pulse of detrital sedimentation related to the Borden delta. The Edwardsville Member demonstrates a classic pattern of silts and clayey silts closer to the source area, with lithologic gradation farther offshore into finer muds with interbedded, fossiliferous carbonates.

In the northeastern part of the study area, the base of the Edwardsville Member (Figure 5) consists of a thin, clay shale unit that is succeeded by a series of overlapping wedges of repeating sediment types that dip to the southwest. The basal clay shale is interpreted as accumulating slowly by pelagic sedimentation and represents the distal deposits of the regressive sequence. With continued regression, silts and fine sands began to reach the study area, and the rate of sedimentation increased.

In the bedded siltstone lithology, the ideal cycle (Figure 7) contains a lower, planar-laminated interval (units 1A and 1B) interpreted to represent the lower part of the upper flow regime conditions. Unit 1B, in some cycles, grades upward, seemingly conformably, into unit 2; where, unit 2 is believed to represent deposition from the waning portion of a turbidity flow. In other places, unit 2 overlies a burrowed surface developed on 1B where it appears to have been deposited by normal marine currents. The absence of bioturbation features in unit 1, except for the upper surface, indicates that this unit was deposited rapidly with no chance for organic reworking. The abundance of bioturbation structures in unit 2 indicates that sedimentation was relatively slow, allowing time for extensive reworking of the sediment. Although bedding thicknesses and paleocurrents are known for the bedded siltstone intervals, the three-dimensional shape is unknown. No evidence of channels or the presence of channel-fill occurrences, such as shown by Kepferle (1972c) for the Kenwood Member of the Borden Formation, was found in the bedded siltstones, nor was there any differentiation noted of turbidite units into proximal and distal parts.

In the bedded siltstone lithology of the Edwardsville Member, directional structures are present that permit a paleocurrent reconstruction for this portion of the Muldraugh Formation (Figure 10). The directional indicators are divided into two groups: furrowcasts and ripple laminations. The furrowcast group of structures is composed of three members which are interrelated: furrow flutecasts, bifurcating furrowcasts, and non-bifurcating furrowcasts. All are characterized by ridges and grooves of shallow relief parallel to the direction of current flow. In the Edwardsville Member, in all instances where bifurcating furrowcasts were observed, the bifurcations open to the northeast. The bifurcations are in the upcurrent direction, determined independently from unidirectional current flow structures in the Kenwood (Kepferle, 1972c) and from the paleoslope, as indicated by isopachs of the Borden interval (Lineback, 1966; Peterson and Kepferle, 1970). The second group of directional structures measured was the strike of the trough or crest of ripple laminae. Ripple laminae are internal structures, as opposed to ripple marks, which are surface forms.

A comparison of the direction of current flow, derived from furrowcast data (S34°W) with that from the ripple laminations (S26°W) shows a very close agreement. Combining both sets of data, S30°W is derived as the mean direction of current flow in the bedded siltstone portion of the Edwardsville Member. Linear directional indicator data are summarized in Table 4.

Siltstones in the Kenwood Member of the Borden Formation were interpreted by Kepferle (1972c) to be deposited by turbidity currents that spread out at the slope base. This fan-shaped pattern is shown (Figure 10) in the gradual swing in paleocurrent indicators from about S80°W in the north to S50°W in the south. Paleocurrent evidence for the Edwardsville Member seems to indicate a more northerly south than the Kenwood Member by the mean current direction moving towards S30°W; however, the area studied is probably not large enough to recognize the overall pattern of paleocurrents during Edwardsville deposition. Just to the north and northwest of the map area are bedded siltstone units within the Edwardsville Member, such as the beds termed the Fordyce Knob Sandstone (Stockdale, 1931), that are similar to those in this study and should contain directional features.

The approximate position of the Borden delta front is shown on Figure 10. The delta front is defined as the area where the Borden undergoes an abrupt decrease in thickness towards the southwest. The delta front, interpreted by Peterson and Kepferle (1970) to be the foreset slope of the delta, is in the shallow subsurface.

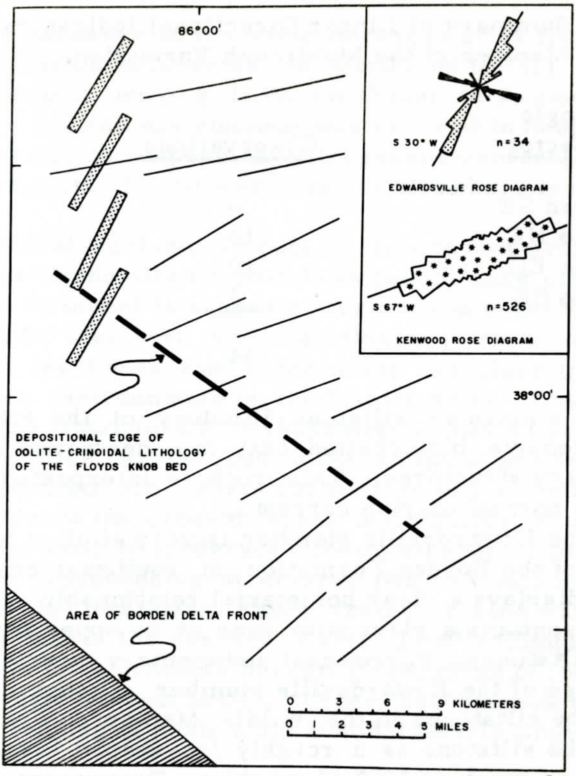

Figure 10. Moving average map comparing paleocurrents in the Edwardsville Member of the Muldraugh Formation (coarse stripple sense of current bars) with those in the Kenwood Member of the Borden Formation (single line sense of current bars). Data on the Kenwood Member are from Kepferle (1972c, Figure 3b). Kepferle's rose diagram includes measurements to the southeast of this map area; however, they are similar to those shown here. On the Edwardsville Member rose diagram, the coarse striple pattern shows distribution of furrowcast structures, the black area shows the distribution of ripple laminae data.

In summation, the bedded siltstone lithology of the Edwardsville Member was produced by periodic storms or major floods which eroded sediment on the upper delta platform. This sediment moved down the paleoslope to the southwest to a position on the outer edge of the delta platform. Some of this material was emplaced by minor turbidity flows and the rest by normal marine currents.

Table 4. Summary of Linear Directional Indicators in the Edwardsville Member of the Muldraugh Formation.

Quadrangle and 1/9 Sector	Observations	Artithmetic Mean
Georgetown SE	6	31°
Lanesville NE	16	30°
Lanesville EC	10	20°
Lanesville SE	2	39°
Total	34	30°

The massive siltstone lithology of the Edwardsville Member shows intensive bioturbation that has destroyed most of the primary sedimentary structures. This rock is interpreted as being deposited slowly by normal marine currents.

The Edwardsville Member is very similar to the Wildie Siltstone Member of the Borden Formation in southeast-central Kentucky, with which it displays a clear homotaxial relationship. The Wildie Siltstone Member contains a glauconite zone at its upper and lower contacts and bedded siltstones. Furrowcast sedimentary structures similar to those on the base of the Edwardsville Member siltstones are present on the base of the siltstones in the Wildie Member. Weir (1970, Figure 13) mapped the siltstone as a roughly fan-shaped body, with the siltstone reaching as much as 10 ft (3 m) thick. Furrowcasts in the Wildie Member are subparallel to the various lobes of the siltstone body. The bedded siltstone intervals in the Edwardsville Member may have an areal pattern similar to those mapped by Weir (1970, Figure 13). Weir (1970) and Weir and Gualtieri (1970) interpreted the siltstone beds as deltaic turbidites similar to those in the Farmers Silstone Member of the Borden Formation in northeastern Kentucky and the Kenwood Siltstone Member of the Borden Formation in north-central Kentucky. Figure 3 in Weir and others (1966) suggests, however, an outer platform depositional site for the siltstone in the Wildie Member and not a base-of-slope depositional environment, as suggested by Kepferle (1972c) for the Kenwood and Farmers Siltstone Members of the Borden Formation.

The depositional relationship between the major subunit of massive siltstones and bedded siltstones in the northeast with the carbonates and shales to the southwest is not well understood, but both are presumed to be contemporaneous. The shaly aspect of these beds is related to their more distal position from the source area to the northeast. The carbonate content is probably related to a depositional environment with a relatively low rate of terrigenous detrital influx. The shale-carbonate subunit thickens southwestward, with both the shale and crinoidal limestone lithologies increasing in thickness down the paleoslope. The increase in the shale thickness away from the source area, determined by paleocurrents, is attributed to: (1) sedimentary bypassing by currents moving from northeast to southwest down the paleoslope until a quieter environment was reached in deeper water; or (2) sediments carried in suspension by longshore currents being derived from an active delta distributary system elsewhere (probably from the north).

A depositional-ecological model similar to that proposed by Lane (1973) for the Crawfordsville Crinoid Beds (which Lane correlated with the Edwardsville Member) is suggested for the shale-carbonate subunit of the Edwardsville Member. Crinoid colonies (with associated biota) grew in clusters, level with the surrounding sea floor in upper slope -edge-of-platform environments on the Borden delta. The distribution of the colonies was random or perhaps determined by areas of higher current intensity which carried food to the animals and prevented fine sediments from settling out. Variation in size of the crinoid colonies through time produced the present areal distribution of the limestone beds. Evidence of weak to moderate current strength is present in these rocks. Although crossbedding structures (which would indicate movement of sediment by migrating ripple trains or sand waves) were not noted in the limestones, the absence of articulated columnals and sharp upper and lower bedding contacts indicates reworking of local patches of crinoids by moderate current action. Other fossiliferous patches show an intimate interbedding of shale and crinoidal debris limestone, indicating weak currents.

Depositional Environments in the Ramp Creek Member

The apparent absence of the slowly accumulated, transgressive, basal Ramp Creek glauconite zone in the southwesternmost part of the study area in Indiana and Kentucky may be due to: (1) continuous sedimentation, or (2) strong currents near the platform-slope break which scoured away the glauconite. After cessation of glauconite formation, dominantly carbonate sands and muds began to mantle the pre-existing depositional topography. In the vicinity of the platform-slope break (essentially this study area), large crinoid meadows flourished. Skeletal remains of these animals were swept both seaward, to form large crinoidal limestone bodies, such as the Cane Valley Limestone Member of the Fort Payne Formation (Sedimentation Seminar, 1972), and shoreward, to form thin crinoid biosparites interbedded with dolomicrites.

Chowns and Elkins (1974) recognized a sabkha environment in the Fort Payne Formation on the basis of dolomites interpreted as replacing micrites and geode pseudomorphs of early diagenetic anhydrite. However, features indicative of subaerial exposure (algal lamination, birdseye structures, dessication features, and evaporation crusts) are missing in the Fort Payne Formation (Chowns and Elkins, 1974) and also in the Ramp Creek Member of this study. Probably, shallow, moderate, and deeper water environments are represented in the Ramp Creek-Fort Payne complex of rocks. Those in the study area represent deposits in a platform environment.

South and southwestward progradation of carbonates generated in a shallow-water environment and swept over the edge of the platform caused infilling of the basin left at the end of clastic deposition. The deposition of the shallow-water Harrodsburg Limestone concluded the infilling process, resulting in a shallow platform sea (Peterson and Kepferle, 1970).

SUMMARY

Event	Remarks
1. Cessation of Borden sedimentation	Reflects lack of major tectonic events in the source region of northeastern North America at the close of the Acadian orogeny; pronounced depositional topography developed at southwestward limit of thick, Borden delta sediments.
2. Formation of lag conglomerates of early diagenetic siltstone concretions derived from erosion of the upper part of the Borden Formation	Lowering of sea level exposing upper part of Borden delta to erosion.
3. Crinoid biosparites and oosparites of Floyds Knob	Shallow water, high energy carbonates formed in initial stage of marine transgression.
4. Deposition of basal Edwardsville glauconite zone	Continued marine transgression produced by important (eustatic?) sea level rise increases distance to sediment supply and allows deposition of glauconite and phosphates.
5. Accumulation of Edwardsville	Regressive deposit with shore moving southwestward showing classic pattern of silts and clayey silts closer to the source area (paleocurrents from N30°E) while grading laterally further offshore (and down the paleoslope) into finer muds with interbedded crinoidal carbonates.
6. Deposition of basal Ramp Creek glauconite zone	Halt in sedimentation owing to (eustatic?) sea level rise with conditions similar to #4 above.
7. Crinoid biosparites and dolomicrites of Ramp Creek	Carbonate muds occurring on platform and in basin environments while crinoid meadows growing at the platform-slope break contribute skeletal debris to both environments.
8. Crinoid biosparites of Harrodsburg	Continued carbonate sedimentation from Ramp Creek time has infilled basin left at end of clastic sedimentation to produce a shallow water, carbonate bank.

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STRATIGRAPHY OF KIAWAH ISLAND BEACH RIDGES

By

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ABSTRACT

Exposures in the cutbanks of tidal creeks on Kiawah Island, South Carolina display particularly well the internal stratification of several Holocene beach ridges. The observed lithologic sequences in the landward and seaward sides of these ridges reflect the translation of subenvironments observable on modern beaches. The landward sequence is: marsh-washover-berm crest/incipient dunes-dunes. The seaward sequence is: marsh-beachface-runnel-berm crest/incipient dunes-dunes. The transgressive or regressive nature of these deposits cannot be determined from small-scale sequences. Rather, the entire lithosome geometry must be considered. This geometry can be expected to vary greatly along depositional strike, largely as a function of the physical processes that control the geomorphology of the beach ridges themselves.

INTRODUCTION

The geologic literature contains many descriptions of barrier-associated sedimentary structures, textures, and facies distributions (see Berryhill et al., 1969; Hayes, 1969; Fisher and Brown, 1972; and Schwartz, 1973 for a summary. Most previous work on the stratigraphy of beach ridges and barrier, however, has emphasized large-scale spatial relationships between barrier sands and adjacent lagoonal and open marine sediments. Notable examples are the studies of Fisk (1959), Byrne et al., (1959) Bernard et al., (1962), Van Straaten (1965), Dillon (1970), Otvos (1970), and Kraft (1971). Smaller scale lithologic relationships within individual recent sand bodies are seldom considered in detail (Hayes, 1969). From the literature on ancient sediments, several exceptions are the works of Sabins (1962), Berg and Davies (1968), Davies et al., (1971), Ferm et al., (1971), Weber (1971), Land (1972), Hobday (1974), and Klein (1974). The purpose of this paper is to describe lithologic sequences observed in beach ridges on Kiawah Island, South Carolina, and to suggest some applications of the sequences

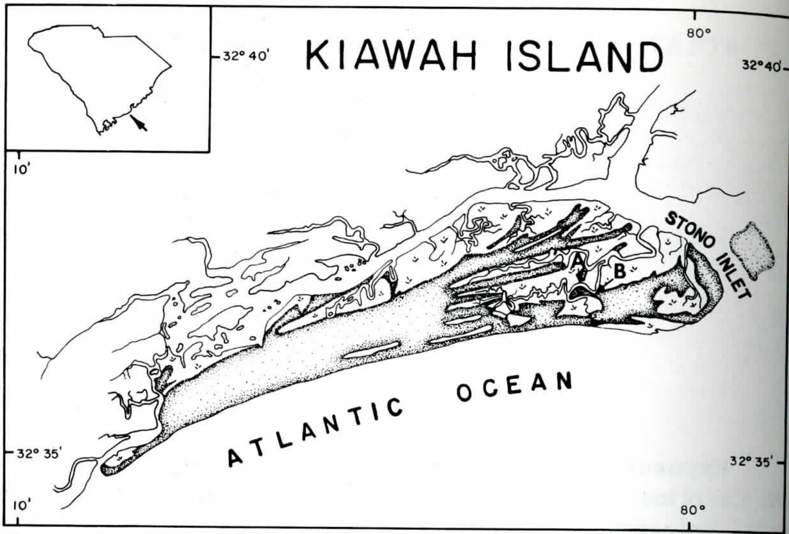


Figure 1. Map showing the distribution of beach ridges along the eastern end of Kiawah Island, S. C. Numerous cutbanks formed by the migration of tidal creeks through these beach ridges provide good exposures of internal ridge structure. The locations of detailed sections (A-landward; B-seaward sequences) are also shown.

in the interpretation of ancient sedimentary environments.

Kiawah Island is characterized by a complex of bifurcating beach ridges of Holocene and Pleistocene age (Hayes et al., 1975). Extensive fields of these ridges are common to the South Carolina and Georgia coasts (Ziegler, 1959; Oertel, 1975). Near the eastern end of Kiawah, inter-ridge swales are veneered with recent salt marsh deposits. In several places, the cutbanks of meandering tidal creeks lacing these marshes have exposed complete sections through beach ridges (Figure 1). Such exposures are not uncommon along the South Carolina coast. The exposures are often quite large and frequently display up to 7 m of section. Sedimentary structures are usually well-preserved, and complete vertical sequences can be studied with a minimum of outcrop preparation.

Acknowledgments

Funds for this project were provided by the Geology Department, University of South Carolina. Ian Fischer, Larry Ward, and Anne Heckel assisted in the field. Miles O. Hayes and John C. Kraft provided helpful interpretive suggestions in the field. Thanks to Dennis Hubbard for manuscript review. Jean Hungiville typed the manuscript.

INTERNAL GEOMETRY

In general, South Carolina beach ridges are composed of clean, well-sorted fine sand. They are underlain by, and interfingered with, organic-rich, silty muds deposited in a salt marsh environment and are capped by thick growths of climax forest. Two representative sections will be described here, one each from the landward and seaward sides of two larger ridges location in the beach-ridge complex at the eastern end of Kiawah Island (Figure 1). The lithologic transition between these two types of sequences has not been observed but is inferred to be similar to the lateral relationships described from trenches in modern berm crests.

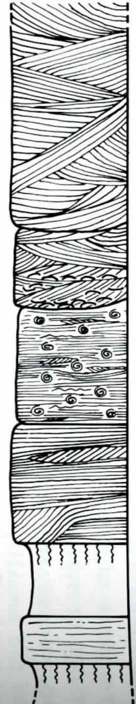
Landward Sequence

This sequence crops out in a cut on the south side of Bass Creek (Figure 1) and has at its base an undetermined thickness of organic-rich rooted muds. Sand facies grade upward from plane beds into small scale trough crossbeds and finally into large scale planar and trough crossbeds (Figure 2). This sequence is interpreted as representing salt marsh-washover-windflats-dunes.

The lowermost salt marsh unit (Figure 2) is composed of organic-rich dark gray silty clay. Remains of Spartina sp. are abundant, both as roots and as stubs protruding into the overlying sand. The upper 1 to 3 mm commonly contains a fine shell hash in which the average grain size is less than 1 mm.

Overlying, and often interfingering with, the salt marsh sediment is a 1 m thick unit of well-sorted fine sand (μ=2.7Φ ; σ = 0.7Φ) interpreted as washover material (Figure 2) on the basis of sedimentary structures described as characteristic of washover sands by Hayes (1967), Andrews (1970), Nordquist (1972), and Schwartz (1975). The unit is dominated by plane beds, but contains occasional 20 cm thick lenticular beds that display internal planar cross-stratification and pinch out within the scale of the outcrop (less than 10 m). The uppermost plane beds contain occasional washed-out ripples, both as thin streaks and as beds up to 10 cm thick. Inclination of the plane beds was too low to measure. Preliminary, but inconclusive, slipface azimuth measurements of the planar crossbeds indicate a mode roughly normal to the ridge trend, dipping in an onshore direction. Heavy mineral laminae accentuate bedding. Biological activity in this unit is high. Tests of the small gastropod Ilyanassa obsoleta, common to the high intertidal marsh, are abundant. The entire unit is bioturbated. Burrows are up to 4 cm in diameter and resemble burrows made by the fiddler crab, Uca sp. In the lower 30 cm of this unit, which is normally stained green by the reducing conditions of the underlying marsh, burrows are filled with lighter colored sand from the upper part of the unit.

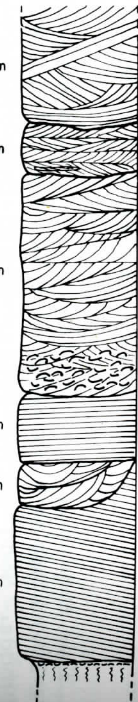
BEACH RIDGE LITHOLOGIES
LANDWARD SEQUENCE



UNIT	ENVIRONMENT	DESCRIPTION
4.7 m	DUNES	Planar and trough cross-beds, sets to 2 m; burrows to 5 cm; rooted.
30 cm	BERM CREST, INCIPIENT DUNES	Trough crossbeds, sets to 20 cm; shell lag.
50 cm	WASHOVER, WIND FLATS	Horizontal plane beds; wind ripples; snails.
50 cm	WASHOVER	Horizontal plane beds; planar crossbeds, lenses to 20 cm; bioturbated.
30 cm	MARSH	Silty clay; H_2S; Spartina roots.
20 cm	WASHOVER	Horizontal plane beds; bioturbated; Fe^{+2} stains.
	MARSH	As above; in-place oyster shells.

Figure 2. Lithologic sequences observed in the landward edge of a Kiawah Island beach ridge.

BEACH RIDGE LITHOLOGIES
SEAWARD SEQUENCE



UNIT	ENVIRONMENT	DESCRIPTION
3.5 m	DUNES	Planar and trough crossbeds, sets to 2 m; burrows to 5 cm; rooted.
25 cm	INCIPIENT DUNES	Ripples; heavy mineral lag deposits.
1.1 m	BERM CREST, INCIPIENT DUNES	Trough crossbeds, sets to 20 cm; shell lag lower 30 cm.
40 cm	BERM CREST	Plane beds, horizontal to gentle landward dip.
20 cm	RUNNEL	Trough crossbedded, lenses to 20 cm.
80 cm	BEACHFACE	Plane beds, dipping 10° seaward.
	MARSH	Silty clay; H_2S; fine shell hash on upper surface; Spartina roots.

Figure 3. Lithologic sequence observed in the seaward edge of a Kiawah Island beach ridge.

Overlying the washover deposits is a 30 cm thick unit of well-sorted fine sand ($\mu = 3.05 \Phi$; $\sigma = 0.6 \Phi$) with trough crossbed sets up to 20 cm thick (Figure 2). Slipfaces are highlighted by heavy mineral laminae. Similar deposits are frequently observed immediately behind the berm crests of washover terraces where well-developed dunes do not occur. This zone of incipient dunes usually represents the transition between the deposition of the washover terrace and the development of wind-shadow dunes. This interpretation is supported by four lines of evidence:

1) by the observation of similar wind modified shell deposits of the crests of modern washovers;
2) by the lateral discontinuity of the unit and the fact that it is gradational into the overlying dune facies;
3) by crossbed slipface azimuths which show no strong mode, but occupy all directions landward of the beach-ridge trend; and
4) by the presence of partially decayed root material, often in abundance, resembling roots of <u>Borrichia frutescens</u> (sea ox-eye). This plant is characteristic of the high marsh environment, and is commonly observed on the margins of washovers, where the bed is modified both by wind and by spring tide flooding.

The sequence is topped by up to 6 m of well-sorted fine sand ($\mu = 2.6 \Phi$; $\sigma = 0.5 \Phi$) deposited in sets of planar and trough crossbeds up to 2 m thick (Figure 2. Slipface azimuths are bimodally distributed and oriented roughly parallel to the beach ridge trend. This cross-bedding trend is common for coastal dunes and represents deposition on wind-shadow dunes (Hayes, pers. comm.). Large burrows (up to 5 cm) are common and are probably left by the ghost crab, <u>Ocypodae albicans.</u>

It is interesting to note that washover and dune sediments have virtually the same grain-size trends. This is probably caused by the fact that the dunes provide the main source for washovers. The relatively finer sand observed in the incipient dune lithology owes its smaller overall grain size largely to high heavy-mineral content, which may approach 90 percent in beds up to 5 cm thick.

Seaward Sequence

This sequence, which represents the aggradation of beach sub-environments, crops out on the north side of Bass Creek (Figure 1). The complete vertical sequence is: marsh-beachface-berm-spring berm runnel/incipient dunes-dunes (Figure 3). Marsh and dune facies are identical to those observed in the landward sequence.

The marsh facies is overlain by up to 1.4 m of well-sorted fine sand ($\mu = 2.5 \Phi$; $\sigma = 0.5 \Phi$) displaying the characteristic beachface plane beds with seaward inclinations of 5 to 10 degrees (Figure 3). Bedding is outlined by heavy mineral laminae, which are often iron-stained. In one section, runnel deposits were observed; these appeared as a lenticular bed containing trough crossbeds and abundant shell material. The trough crossbeds are internally ripple cross-laminated and dip in a general alongshore direction. Analogous structures occurring in runnels on modern beaches usually represent deposits left by cuspate megaripples migrating alongshore in a runnel. At lower tide stages, they are frequently covered by smaller scale bedforms (linguoid ripples and washed-out ripples) caused by late stage ebb flow. No ridge deposits (landward-oriented planar crossbeds) were observed; therefore, the bedding plane surface immediately above the presumed runnel deposits must, of necessity, be erosional.

Overlying the beachface unit is a 1.5 m thick unit of well-sorted fine sand ($\mu = 2.5\,\Phi$; $\sigma = 0.5\,\Phi$) also containing heavy mineral laminae and shell debris. This unit grades upward from bimodal trough crossbeds at the base to ripple cross-laminae at the top. These sands are interpreted as alongshore-migrating megaripples caused by spring tide berm-overtopping, which were subsequently wind-reworked. Currents flowing alongshore in spring berm runnels are a common occurrence (Hayes, 1969). Velocities can exceed 1 m/sec, as observed on Fire Island, New York (John Sanders, pers. comm.).

Spring runnel deposits are overlain by incipient dunes and dunes, the structures and textures of which are as previously described.

Transition Zone

Although not exposed in Kiawah beach ridges, the depositional transition between the washover and beach-associated lithologies is probably similar to that observed in a trench cut through a beach, berm, and washover terrace on Cape Romain (Figure 4). In this cut, beds deposited on the washover and the beach interfinger at the crest of the washover terrace. A few thin, plane beds are continuous over the crest and are conformable to both the landward and seaward dipping slopes. Washover structures include both planar foresets and plane beds, as noted by Schwartz (1975). Beach deposits include ridge-associated landward-dipping planar crossbeds in sets up to 20 cm thick, 5 to 10 degree seaward-dipping plane beds of the swash zone, and a 30 cm thick coarse shell deposit, probably of storm origin.

STRATIGRAPHIC IMPLICATIONS

Since Johnson's (1919) original thoughts on beach ridges and barriers, the origins and stratigraphies of barriers, beach ridges, and cheniers have been argued extensively in the literature. Although a review of these arguments is beyond the scope of this paper (the interested reader should consult Schwartz, 1971, 1973), certain general stratigraphic aspects should be considered prior to the application of beach

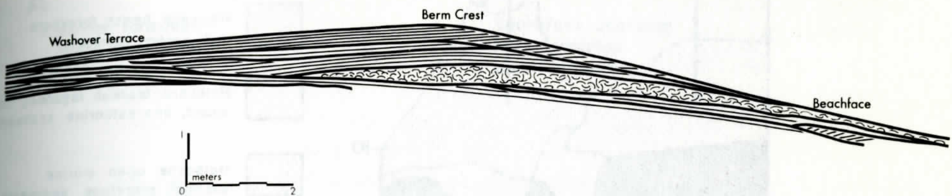


Figure 4. Trench through the crest of a washover terrace on Cape Romain, South Carolina. This trench is presented here to illustrate the probable transition between the Kiawah Island and seaward beach-ridge sequences.

ridge lithologic sequences to paleogeographic reconstruction.

Hoyt (1969) maintained that cheniers and barriers could be distinguished by the shape and extent of the deposits. He argued:

1) Cheniers form from the reworking of a progradational shoreline in areas of high sediment supply,
2) Chenier sand bodies are relatively thin (less than 6 m) and overlie marine sediments,
3) Barrier islands form as the result of lowland flooding in areas behind beach ridges during periods of rising sea level,
4) Barrier sand bodies are relatively thicker (to 47 m) and predate their associated marsh sediments.

The sequences observed on Kiawah Island indicate that such an explanation is too restrictive and favor the broader interpretation of Fisher and Brown (1972):

"Strand plains are prograding coastal areas. This progradation is due to sediment's being transported along the strand by longshore currents. Barrier islands are distinguished from strand plains primarily by the presence of lagoonal facies. It should be stressed that a complete gradation exists from high-destructional deltas through strand plains to barrier islands. As such, it is impossible to distinguish between these three models on the basis of such criteria as vertical sequences. Geometry of the framework sands and the associated facies would be necessary to separate these sytems."

As Hayes, et al., (1969) pointed out, paleogeographic reconstruction based on barrier island sedimentary structures is difficult. These difficulties may not be overcome if reconstructions are based on small-scale lithologic sequences. Depending on the scale of the outcrop examined, completely opposing interpretations could be made about the transgressive or regressive nature of a shoreline. This is illustrated by the sequence of lithologies in the Padre Island, Texas, barrier

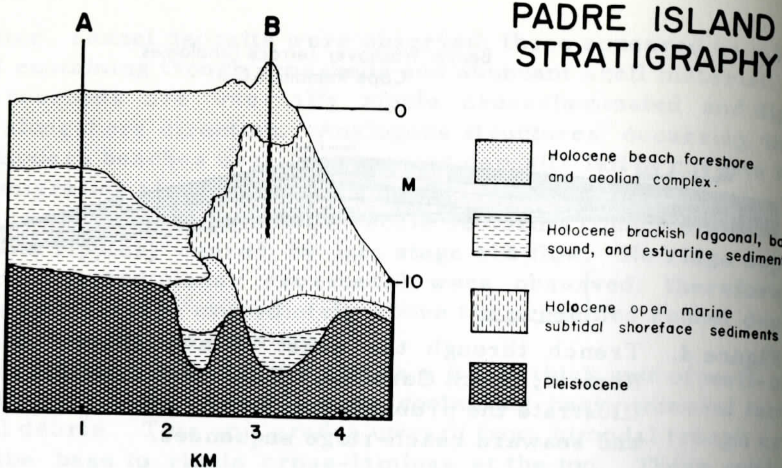


Figure 5. Generalized stratigraphy of Padre Island, Texas after Fisk (1959, Figure 13) and Otvos (1970, Figure 1). Note that an observation of section A only would indicate marine transgression, while an observation of section B only would indicate marine regression.

system (Figure 5). Sections drawn through limited portions of the total geometry indicate opposing trends in shoreline migration. Only when the overall geometry is considered does the shoreline development become clear.

The Kiawah Island shoreline has obviously displayed a period of progradation by beach-ridge barriers similar to that observed by Oertel (1975). Dating has shown most accretion to be younger than 1200 years (Stapor and Mathews, 1976). This ridge development gives the island chenier plain affinities, irrespective of the fact that the ridges did not form by the winnowing of finer-grained sediments deposited during periods of high sediment availability and associated shoreline retreat.

Nevertheless, Kiawah Island, when considered in its entirety, is a barrier that protects lagoonal and marsh facies from open wave attack. Although the barrier island must have predated the development of the original marsh, portions of the barrier lithosome (i. e., beach ridges) do not lend themselves to this simple relative age relationship (Figure 6). Some beach ridges migrated landward, by overwash, across the back-barrier marsh surface during periods of shoreline reorientation. Marsh sediments were then eroded on the beachface, a process observable on Kiawah beaches today. This phenomenon was discussed in detail by Fischer (1961) and Swift (1968) with regard to New Jersey beaches and the importance of this process to the rock record, respectively. The accretion of a new ridge provided an inter-ridge swale in

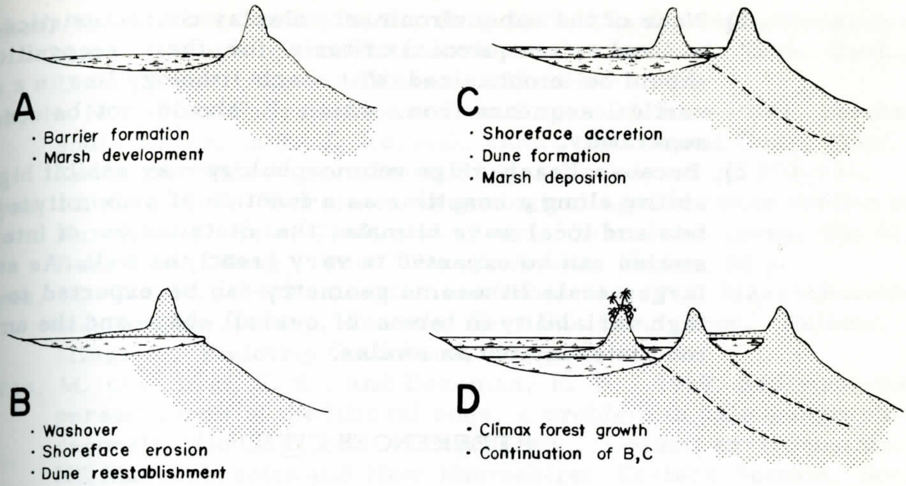


Figure 6. Proposed development of the Kiawah Island barrier - beach ridge lithosome. Note that, in C, recent marsh sediments deposited in inter-ridge swales are contiguous with underlying older marsh sediments. In D, marsh sediments comprise a lenticular body enclosed in washover and beach ridge sands.

which marsh sediments were deposited. Short cores in these swales, along with the outcrop relationships displayed along the tidal creek banks, show that this younger marsh veneer is often continuous with older marsh cropping out on the front of the previous ridge (Figure 6). This could ultimately encapsulate the beach ridge sand body in finer-grained sediments, as an entity separate from the main barrier lithosome. In some cases, washover sands connect the younger ridge with the previous ridge (Figure 6).

The shoreface erosion process poses interesting problems for the dating of beach ridges by the use of shell material. Because older shells from the back-barrier marsh sediments are continously reworked on the beach (Fischer, 1961) they stand a good chance of being washed over the berm and incorporated into deposits that will ultimately be preserved as part of a beach ridge (Carter, 1976). Indeed, shell material observed in the Kiawah beach ridges appeared to originate from two distinct populations: one of fairly fresh unweathered shells, and one of blackened, heavily pitted shells similar to those being eroded from marsh sediments in the beachface.

SUMMARY

Although beach-ridge subenvironments should certainly be recognizable in the rock record, two points should be made with respect to their recognition and interpretation:

1) None of the subenvironments display characteristics that can be used as unequivocal criteria for their recognition. It should be emphasized that each lithology lies in a distinct vertical sequence from which it should not be considered separately.

2) Because beach-ridge geomorphology may exhibit high variability along a coastline as a function of proximity to tidal inlets and local wave climate, the distribution of inter-ridge swales can be expected to vary greatly as well. As such, the larger scale lithosome geometry can be expected to display high variability in terms of overall shape and the amount of mud intercalated as swales.

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GRAIN SIZE AND HEAVY MINERAL ANALYSES OF THE WILCOX AND CLAIBORNE FORMATIONS IN HENRY, WEAKLEY, AND CARROLL COUNTIES, TENNESSEE

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ABSTRACT

Grain size and heavy mineral analyses, included as part of the petrologic investigation of the Wilcox (Lower Eocene) and Claiborne (Middle Eocene) Formations in a 900 square mile area of northwestern Tennessee, indicate that they are best differentiated mineralogically. Methods of investigation included: grain size analysis of 116 sand samples on a quarter phi interval basis; multivariate discriminant analysis of the textural parameters, mean grain size, sorting, skewness and kurtosis; and statistical analysis of the heavy mineralogy of 52 samples.

The Wilcox and Claiborne Formations are moderately well sorted, fine-grained sands that contain discontinuous clay lenses at various horizons within the formations. Multivariate discriminant analysis of the textural parameters indicates the formations cannot be differentiated with a high level of confidence.

Differences in accessory mineralogy permit differentiation of the formations with a high level of confidence. While each formation has the same heavy mineral assemblage, namely ilmenite, leucoxene, zircon, kyanite, staurolite, tourmaline and rutile, Wilcox heavy mineral suites contain more ilmenite but less leucoxene and zircon than Claiborne suites. Also, beds of "sawdust sand" occur only in the Wilcox Formation.

INTRODUCTION

The Wilcox (Lower Eocene) and Claiborne (Middle Eocene) Formations in northwestern Tennessee are lithologically similar and, for the most part, lack distinguishing characteristics, making it very difficult to distinguish them in the field. Consequently, as part of the petrologic investigation of these formations, quantitative criteria were used to determine textural and mineralogical bases for differentiating them.

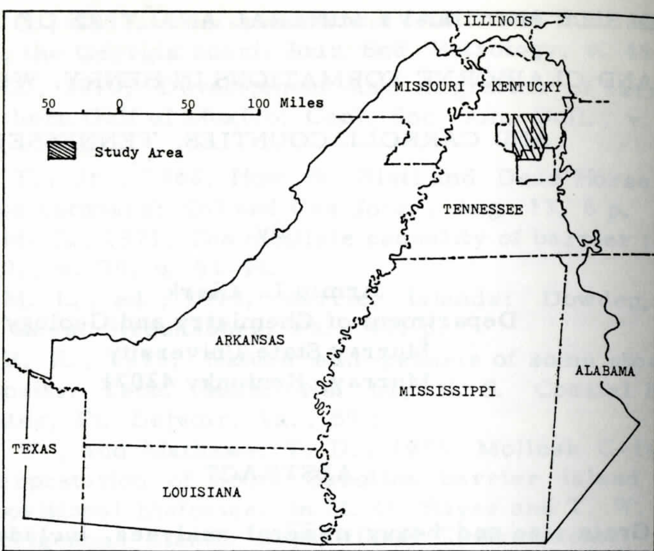

Figure 1. Location map of study area.

Acknowledgments

This study is based on a Ph. D. dissertation completed at the University of Tennessee. The writer expresses appreciation to Lawrence T. Larson, who served as his adviser. Appreciation is also expressed to William S. Parks of the U. S. Geological Survey for his generous help.

GEOLOGIC SETTING

The study area (Figure 1) extends north-south about 35 miles and east-west about 25 miles and includes parts of Henry, Weakley and Carroll Counties of western Tennessee. It is located in the northeastern part of the Mississippi embayment, a wedge-shaped region extending from southern Illinois southward to about the 32° N parallel. Geologically, the Mississippi embayment is a southward-plunging syncline whose axis roughly follows the Mississippi River. The study area is on the east limb of this syncline. The Wilcox and Claiborne Formations strike approximately north-south and dip about 30 feet per mile to the west.

STRATIGRAPHY

Varicolored and heterogeneous, the Wilcox and Claiborne in the study area are nonmarine formations that consist chiefly of sand, silt and clay. These sediments are interbedded and interlensed and no lithology or sequence of lithologies is laterally persistent for any great distance. The sand, chiefly quartz, is sparsely to very micaceous. Bedding varies from thick to very thin and is horizontal and irregular, crossbedded and lenticular. "Sawdust sand" (Whitlatch, 1940), a fine, clayey, micaceous, white specked sand, is a distinctive lithology found only in widely scattered outcrops, 5 to 10 feet thick, in the basal part of the Wilcox Formation in the easternmost part of the study area. Clay beds of the formations range from a fraction of an inch to several tens of feet thick and usually occur in discontinuous lenses that apparently are not related to a particular stratigraphic position within the formations. Lignite beds, which vary in thickness from several inches to several feet, separate some clay lenses from the overlying sands. Leaf fossils and lenses of lignite are present within some clay beds.

GRAIN SIZE ANALYSIS

Sand samples were collected from road cuts and overburden faces in clay pits. The sampling techniques employed are those outlined by Folk (1965) as to sample size and method of collection. Each sample was tagged, numbered and its location and tag number recorded on the topographic map. One hundred and sixteen samples from 57 sites were obtained. Sample sites were listed and their locations given in Tennessee Rectangular Coordinates (Clark, 1973). Each sample site on the topographic map was identified as either within the Wilcox or Claiborne Formation and its vertical location above the base of the formation estimated (Clark, 1973) by William S. Parks of the U. S. Geological Survey, on the basis of his geologic mapping in the area (unpublished at the time of this study). Stratigraphic control for the U. S. Geological Survey maps was by means of four drill holes. Of the total 116 samples from 57 sites, 23 samples at 16 sites were estimated to have been taken from the Wilcox Formation and 93 samples at 41 sites were judged to have been taken from the Claiborne Formation.

Grain size distribution for each sample was determined by standard sedimentological procedures (Folk, 1965) using quarter phi sieve intervals. The weight percent of the finer than 4. 5 phi fraction (pan fraction) exceeded 5 percent in 15 samples and grain sizes of these pan fractions were measured by pipette analysis. Graphic analysis of each sample, following the technique of Folk (1965), was completed and the percentiles used in calculating the values of the textural parameters Graphic Mean (M_Z), Inclusive Graphic Standard Deviation(σ_I), Inclusive Graphic Skewness (Sk_I), and Graphic Kurtosis (K_G) for all samples (Clark, 1973).

Grain size analyses for both formations are summarized and compared in Table 1. Comparison of both mean sorting values with the sorting scale suggests the Claiborne Formation is better sorted than

Table 1. Comparison of Grain Size Analyses of the Wilcox and Claiborne Formations.

Graphic Grain Size Parameter	Total Range	Average Value	Standard Deviation	95 Percent Confidence	Verbal Designation
Mean Size (M_z)					
Claiborne	0.90-4.32 phi	2.13 phi	.79 phi	.16 phi	fine sand
Wilcox	0.90-3.76 phi	2.11 phi	.60 phi	.26 phi	fine sand
Sorting (σ_I)					
Claiborne	0.25-1.70 phi	0.59 phi	.22 phi	.05 phi	moderately well sorted
Wilcox	0.32-1.74 phi	0.71 phi	.31 phi	.13 phi	moderately well sorted
Skewness (Sk_I)					
Claiborne	-0.60 to +0.53	+0.10	.20	.04	fine skewed
Wilcox	-0.41 to +0.59	+0.19	.22	.10	fine skewed
Kurtosis (K_G)					
Claiborne	0.80-2.47	1.28	.30	.06	leptokurtic
Wilcox	0.97-2.06	1.30	.32	.14	leptokurtic

the Wilcox Formation. Comparison of mean skewness values with the skewness scale suggests Wilcox sands are more finely skewed (more silty) than Claiborne sands. There are no significant differences in mean grain size and kurtosis.

MULTIVARIATE DISCRIMINANT ANALYSIS

A multivariate discriminant analysis computer program (Dixon, 1964) was used to differentiate the Wilcox and Claiborne Formations sedimentologically, based on the mapping by Parks (Fig. 2, line 1). Sand samples were divided into two groups, one of 23 Wilcox samples from 16 sites and the other of 93 Claiborne samples from 41 sites (Table 2). Four variables (mean grain size, sorting, skewness and kurtosis) from the Wilcox samples were compared with the same four variables from the Claiborne samples. As few as two variables may be compared in this type of analysis, but utilization of four variables increases the degree of discrimination and produces a higher level of confidence in the results. The discriminant function generated by this analysis was:

$$D(M_z, \sigma_I, Sk_I, K_G') = -0.4284M_z + 2.1438\,\sigma_I + 2.4499Sk_I - 4.0633K_G'$$

The discriminant index calculated was -1.4243. Utilizing this discriminant function and discriminant index, separation of the samples into two groups was completed. This classification yielded a differentiation with a 34 percent overlap, samples which on a multivariate discriminant analysis basis were incorrectly classified as to formation. Using only the two variables which showed the greatest differences between their group means, sorting and skewness, the overlap increased to 36 percent. This discriminant analysis suggests that the sediments

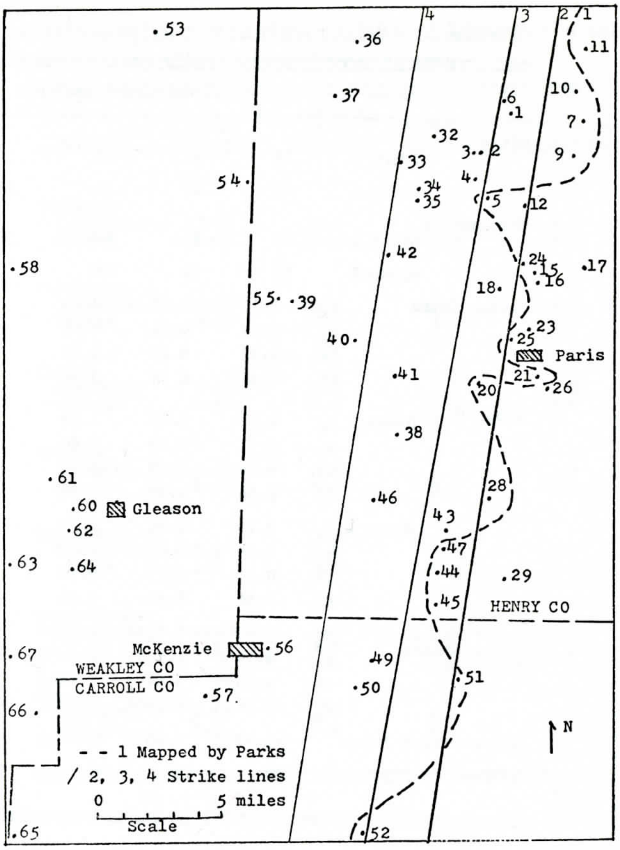

Figure 2. Dividing lines for sample sites for discriminant analyses 1-4.

cannot be differentiated sedimentologically as mapped with a high level of confidence.

The possibility that the sediments could be differentiated sedimentologically, but not as mapped, was also considered. Three additional discriminant analyses (Table 2) were used to test this possibility and to test the possibility that any nearly random strike like would be as valid a sedimentological formational contact. For each analysis separation of the samples into two groups was achieved by a line drawn to follow the general strike of the sediments and moved west of the previous position for each successive analysis (Figure 2). Utilizing the discriminant function and discriminant index for each analysis, separation of the samples into two groups was completed. These classifications yielded differentiations with overlaps of 40 percent, 36 percent and 30 percent, respectively. Results of these discriminant analyses are not significantly better or worse than the result obtained in the first discriminant analysis.

Table 2. Discriminant Analyses Data.

			Discriminant Analysis 1	2	3	4
Sites	Group 1		16	16	26	39
	Group 2		41	41	31	18
Samples	Group 1		23	24	42	73
	Group 2		93	92	74	43
Common Means		M_z	2.13	2.13	2.13	2.13
		σ_I	0.61	0.61	0.61	0.61
		Sk_I	0.12	0.12	0.12	0.12
		K'_G	0.56	0.56	0.56	0.56
Group Means	Group 1	M_z	2.11	1.88	1.94	1.89
		σ_I	0.71	0.68	0.64	0.59
		Sk_I	0.19	0.10	0.15	0.11
		K'_G	0.56	0.57	0.56	0.56
	Group 2	M_z	2.13	2.20	2.24	2.55
		σ_I	0.59	0.60	0.60	0.66
		Sk_I	0.10	0.13	0.10	0.15
		K'_G	0.56	0.56	0.56	0.57
Coefficients in Discriminant Functions		M_z	-0.4284	-0.7390	-0.9388	-1.5675
		σ_I	2.1438	1.5365	1.0259	-0.7742
		Sk_I	2.4499	0.0433	2.4445	1.4431
		K'_G	-4.0633	4.2937	-1.5211	-0.6663
Discriminant Index			-1.4243	1.8973	-1.8614	-4.1494
Percent Overlap			34	40	36	30

In summary, while comparison of parameters in Table 1 suggests Claiborne sands are better sorted and less silty than Wilcox sands and thus may be differentiated, results obtained in the multivariate discriminant analysis indicate the sediments cannot be differentiated with a high level of confidence.

HEAVY MINERAL ANALYSIS

The samples from which grain size analysis data were derived were used for heavy mineral analysis. Heavy minerals were studied from all sand sample sites except five which duplicate the stratigraphic position of other studied samples. A split of medium, fine and very fine sand (1.5-4.0 phi) was obtained with a microsplitter and the heavy minerals separated from the light minerals using bromoform (S. G. = 2.89), following the procedure outlined in Krumbein and Pettijohn (1938). Heavy minerals representing 52 sample sites were mounted on slides in Canada balsam and examined with the petrographic microscope. At random locations on each slide all grains in the field of view were

Table 3. Summary and Comparison of Heavy Mineral Analysis Data of the Wilcox and Claiborne Formations.

Mineral	Range (Percent)	Average Value (Percent)	Standard Deviation	95 Percent Confidence
Ilmenite				
Claiborne	13-64	38.1	13.3	4.3
Wilcox	38-66	54.6	8.3	5.2
Leucoxene				
Claiborne	3-53	22.6	13.4	4.3
Wilcox	2-20	10.2	5.7	3.6
Zircon				
Claiborne	6-32	21.2	6.7	2.2
Wilcox	4-40	16.3	10.2	7.0
Kyanite				
Claiborne	3-26	8.7	4.4	1.4
Wilcox	1-19	8.8	5.7	3.6
Tourmaline				
Claiborne	2-9	4.8	1.7	0.6
Wilcox	1-9	5.8	2.5	1.6
Staurolite				
Claiborne	1-7	3.0	1.6	0.5
Wilcox	1-7	2.8	2.0	1.2
Rutile				
Claiborne	1-4	1.8	0.8	0.3
Wilcox	1-4	1.9	1.0	0.6

identified and counted. Location of each field of view was determined by a mechanical stage and recorded in order to prevent overlap when a new location was selected. Enough fields of view were used to obtain a count of approximately 300 grains.

Heavy minerals identified include ilmenite, leucoxene, zircon, tourmaline, kyanite, staurolite and rutile. Identification of ilmenite, leucoxene and zircon was confirmed by X-ray powder diffraction. A few long, fibrous grain fragments on several of the slides appeared to be sillimanite but its presence was not confirmed. Heavy mineral separations were tested with a strong magnet before mounting for the presence of magnetite, but none was found.

Relative abundances of the heavy minerals were listed by sample site (Clark, 1973) and are summarized and compared in Table 3. The greatest differences occur in the means of ilmenite, leucoxene and zircon. Comparisons of these means using Student's "t" test indicate they are significantly different (Table 4). The normal deviate function was used to determine the level or degree of differentiation (the value at which no overlap occurs and the percentage of the total samples for which the value holds true).

Table 4. Comparison of Heavy Mineral Means for Wilcox (Tw) and Claiborne (Tc) Formations Using Student's "T" Test and Normal Deviate Function.

	1 Ilmenite		2 Leucoxene		3 Zircon
Calculated "T" Test Value	4.00		3.00		1.90
Significance Level of "T" Value	99%		99%		94%
Dividing Values	Tw >48% Tc <48%		<14% >14%		<19% >19%
Tw Values - no overlap	10		9		9
Tw Values - overlap	2		3		3
Tc Values - no overlap	33		28		26
Tc Values - overlap	7		12		14
Total - no overlap	43		37		35
Percent - no overlap	83%		71%		67%
Normal Deviate Level	80%		75%		60%
	1+2	1+3	2+3		1+2+3
Tw Values - no overlap	8	9	6		7
Tw Values - overlap	0	1	1		0
Tw Values - unclassified	4	2	5		5
Tc Values - no overlap	26	20	15		16
Tc Values - overlap	5	3	2		1
Tc Values - unclassified	9	17	23		23
Totals - no overlap	34	29	21		23
Percent - no overlap	87%	90%	88%		96%
Calculated Level of Differentiation - no overlap	95%	92%	90%		98%

Comparison of ilmenite means indicates a very significant difference between the Wilcox and Claiborne. The number separating nonoverlapping values (48 percent) and the highest level of differentiation (80 percent) were calculated and are given in Table 4. Sample values were compared with the calculated normal deviate value and tabulated. Eighty-three percent of the samples did not overlap. Comparison of the leucoxene means also indicates a very significant difference between the formations. Fourteen percent is the figure that separates nonoverlapping values. The highest level calculated at which no overlap of values occurs is 75 percent. Seventy-one percent of the samples did not overlap. Zircon is the only nonopaque mineral to show a significant difference in means. Nineteen percent is the figure separating nonoverlapping values. The highest level calculated at which no overlap of values occurs is 60 percent. Sixty-seven percent of the samples did not overlap.

When all three heavy minerals are used, one can differentiate the Wilcox and Claiborne Formations with a very high level of confidence. The calculated level is 98 percent and the level for the samples is 96 percent. A high level of differentiation is also possible when using any two of the three minerals. Inspection of Table 4 shows the calculated level of differentiation in all cases is 90 percent or greater while the samples show a differentiation level of approximately 90 percent. Ilmenite is probably the only heavy mineral that could be used singly to differentiate the Wilcox and Claiborne Formations.

CONCLUSIONS

There are two significant mineralogical differences between the Wilcox and Claiborne Formations: (1) The ilmenite-leucoxene-zircon suites differ, that is, Wilcox suites contain more ilmenite and less leucoxene and zircon than Claiborne suites; (2) "Sawdust sand" is present only in the Wilcox Formation. Investigation of textural parameters by multivariate discriminant analysis suggests that they cannot be differentiated with a high level of confidence with the parameters examined.

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THE CHATTAHOOCHEE EMBAYMENT:

DISCUSSION AND REPLY

Discussion

By

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May's (1977) paper in which he redefines the "Chattahoochee Embayment" serves a useful purpose in that it attracts further attention to problems with names and to the existence of proposed structural features in southwestern Georgia and part of Florida. The problems are even greater than he outlined, because he ignored the proposed faults and even a graben in this region. With even a cursory search of the literature, one can find proposed and formally named anticlines where other geologists have shown synclines to be present. One can also find a proposed feature having an axial orientation in virtually any direction he would like.

The principal reason for this discussion is to reply to May's criticism of Patterson and Herrick (1971) for not explaining why we used the term "Apalachicola Embayment" instead of "Chattahoochee Embayment," which May favors. May also stated that the shortening of the name "Gulf Trough of Georgia" to "Gulf Trough" is improper. As we applied the term "Gulf Trough" to most, if not all, of the area that May considers to be the Chattahoochee Embayment, comment on this point will also be made.

Patterson and Herrick were attempting to call attention to the problems, to warn against the acceptance of some questionable proposed structural features, and not to add to the confusion. We intentionally used names in vogue with the Geological Surveys of Georgia and Florida at the time of the writing, and the manuscript was reviewed by both State Geologists before publication. The name "Chattahoochee" for the embayment was not in use at that time by either Survey for what seemed to be very good reasons. To have used this term would have required redefinition (which May has done), and we thought it would do little to clear the confusion.

The reason the term "Apalachicola" was used for a sediment-filled basin of limited extent was because this term was used for the basin in a publication of the Florida Geological Survey (Puri and Vernon, 1964, fig. 2), and the location and extent of the basin were shown in an illustration. The term is from the geographic name for the bay. In the original proposal of the name "Chattahoochee" for the structural feature (Johnson, 1891), the embayment was not illustrated; it was also inadequately described, probably because so little was known about it.

We chose not to redefine the Chattahoochee Embayment for several reasons. First, the name "Chattahoochee" was originally taken from the name of the river flowing into Apalachee (now Apalachicola) Bay, a river which, in 1891, was called the Chattahoochee. Because of the whims of geographic namers, a river of this name has not flowed near the structural feature since 1931. In that year, the Board of Geographic Names terminated the application of the name "Chattahoochee" to this river below its junction with the Flint River in Georgia. All quality maps in atlases and those published by the Federal and State governments apply the name "Apalachicola" to the part of the river that is in Florida. Secondly, the proper name (grammatical usage) "Chattahoochee" has been applied to several questionable inferred anticlines, and the use of the same name for both an anticline and an embayment seemed undesirable. Thirdly, the feature under discussion seemed to us to be shaped more like a trough than an embayment. Its areal extent and orientation are also markedly different from those originally and inadequately described.

May's paper stated rather clearly that the name "Gulf Trough of Georgia," as it appeared in publications of the Georgia Geological Survey, has been improperly changed to the "Gulf Trough," as used in a Florida Geological Survey publication. He has a point, because the shorter name leaves out the geographic location. This brings up the problem of how to correct the difficulty. Would it be better to refer to the structural feature as the "Gulf Trough of Georgia," as simply the "Gulf Trough," the "Chattahoochee Embayment" (using a misleading unrelated geographic term), or as some other name that may have appeared in print? None of these names are above criticism but it would seem that any re-redefinition at this time would only contribute to the confusion. Probably the best course of action now is for geologists to bear in mind that the names problem exists and to wait until a good deal more geologic data on this structural feature and its origin are available before attempting to unravel the mess.

Another point worth some thought is, how does one determine what is proper and what is not in naming a structural feature? Possibly Patterson and Herrick as well as May were a bit improper in stating that usages preferred by others are not proper. So far as I can determine, no codes or ground rules for naming such features have been published. Perhaps some knowledgeable and interested geologist--and May may be a likely candidate--should draft a code of procedure and publish arguments for its acceptance in a geological journal having nationwide distribution.

Reply

By

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In general terms, Patterson (1978) and I (May, 1977) agree on the existence of a problem in nomenclature with regard to the "Chattahoochee Embayment". We, further, agree that there is a need for guidelines to govern the naming of structural features such as this. I would go even farther in suggesting that rules of nomenclature are needed for guidance in the naming of any geologic feature that requires a name. Do not structural geologists and geomorphologists share with stratigraphers and paleotologists the need for accurate communication?

Our area of agreement, however, ends at this point. A number of his comments require a response. I list these below in the order in which they occur in his discussion:

1. It was not my intent to "redefine" the Chattahoochee Embayment. To the contrary, I recommended discounting the numerous informal "redefinitions" that have appeared in the literature since Johnson's original definition (1891).

2. I acknowledge that I did not discuss the "proposed faults and even a graben" that have been reported. The omission was for two reasons: (a) there is little agreement as to the location, orientation, and even existence of these faults; and (b) it was not my purpose to discuss the many problems that exist in the geologic interpretation of the feature, but rather to point out the difficulties in nomenclature that have arisen. This purpose was explicitly stated in the second paragraph of my paper (page 149). I cited 26 pertinent publications, and reviewed (but did not cite) a number of additional papers. There were certainly a few that I may have missed, but my review was not "cursory".

3. With regard to my reference to the term "Gulf Trough of Georgia" (Herrick and Vorhis, 1963) and its shortening to "Gulf Trough" (Sever, 1966), my exact wording was that it was "unfortunate" (May, 1977) not "improper". It was "unfortunate" because of the similarity to, and possible confusion with, the term "Gulf Coast Geosyncline". The term "Gulf Trough of Georgia" may be proper if its usage is restricted to the Georgia portion of the larger feature. However, Herrick and Vorhis (1963, p. 55) formally and specifically proposed the name "Gulf Trough of Georgia" for a feature located in southwest Georgia, but described by Applin and Applin, (1944, p. 1727) as "extending southwestward across Georgia through the Tallahassee area of Florida to the Gulf of Mexico." There was no clearly stated effort to restrict usage to the Georgia portion, hence the term was inappropriate, as well as improper.

4. I am not certain that the fact that a name that is temporarily "in vogue" is valid justification for its use. My paper was also reviewed by the State Geologist and geologic staff of the Florida Survey. No one advised me that the term "Apalachicola Embayment" was preferred ...though it had been used in recent publications. Puri and Vernon's publication and illustration (1964, fig. 2) does not lend strong support to the argument either. The figure is difficult to interpret. It appears to show the Suwannee Straits where the Chattahoochee Embayment ought to be and the Apalachicola Embayment some distance to the southwest. They do not discuss the Apalachicola Embayment at all in the text, but in their discussion of the Suwannee Straits they refer to the above quoted Applin and Applin (1944) description. Again there is confusion as to nomenclature.

I erroneously attributed the introduction of the term "Apalachicola Embayment" to Toulmin (1955). It was apparently introduced by Pressler (1947) in a paper that is remarkable in that the geology of Florida and parts of the adjacent states is discussed, yet only two references are cited.

5. Johnson's original description was quite explicit for its time...that is, in light of the absence of geologic data upon which to make an interpretation. In addition to describing specifically the boundaries of the embayment as it was thought to be in Miocene time, he likened it to the modern Apalachee Bay, which brings me to my next point.

6. Patterson has confused Apalachicola Bay with Apalachee Bay. Even a cursory viewing of U. S. G. S. topographic maps (either 1:250,000 or 1:500,000 scale) clearly reveals the position of these bays. They are not the same. Apalachee Bay, to which Johnson and I referred, is formed by a large embayment of the Gulf of Mexico south of Tallahassee. Apalachicola Bay, which has nothing to do with the present discussion, is a estuary/lagoon complex further west near the town of Apalachicola. Patterson's historical recounting of the renaming of the lower Chattahoochee River to the Apalachicola River is irrelevant. It was precisely this type of error and confusion that prompted me to write the original paper.

7. I see no fault in applying the name "Chattahoochee" to the adjacent positive structural feature (that has been called variously an anticline, an arch, and (most recently) an uplift (Price and Whetstone, 1977)). The Chattahoochee "Uplift" and "Embayment" are geologically complementary features and the correct interpretation of one must account for the other.

8. The Georgia portion of the feature is a trough, but the larger Florida portion is clearly an embayment with structure contours that open broadly to the Gulf of Mexico (May, 1977, p. 150). It is reminiscent of Apalachee Bay, except that it was connected through a northeastward extension to the Atlantic Ocean.

9. I would consider the term "Chattahoochee" to be no more

"a misleading geographic term", than is "Gulf Trough of Georgia", "Gulf Trough", etc. I do not argue that "Chattahoochee Embayment" is the best name (I, personally, favor Johnson's original inclination toward "Apalachee Embayment"), but that it has priority and, in the absence of a valid reason to change it, it should be used.

10. I cannot agree that the best course of action is to acknowledge, but ignore, the problem. The problem involves much more than just the Chattahoochee Embayment. I fail to see the logic in delaying the settling of a proper name for a feature until we have completely unraveled its geologic history. In the meantime, what do we call it?

Finally, how are we to face this problem? I suggest that priority based on seniority be the guide and that the local state geologic agency be responsible for establishing what name has priority. If this task puts an unacceptable additional burden on the state agency, perhaps a commission of interested geologists residing within the state could perform it. Where a geologic feature (structural or geomorphic) involves more than one state, they should coordinate on the naming (if they cannot agree, perhaps the United States Geological Survey could act as arbiter). The U. S. G. S. publishes lists of stratigraphic names used in the states. Why could they not also publish names of other geologic units found within states so that out-of-state geologists will know what term is used?

If the state commission decides that for good reason the first-used name is not appropriate, they should formally redefine through publication and the U. S. G. S. should accept and publish the new name. Redefinition should be kept to a minimum, however, since previous publications using the former name cannot be changed.

This is not nearly as sticky a problem as that involving stratigraphy and problems of correlation or paleontology and problems of taxonomy. It seems to me that with moderate effort, care, and common sense these problems could be avoided. I hope that we do not have to resort to an elaborate Code or to National and International Commissions, etc. in order to keep our names straight.

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